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Scientific Report No. 212

RECOMBINATION AND TRANSPORT IN THE NIGHTTIME F REGION OF THE IONOSPHERE

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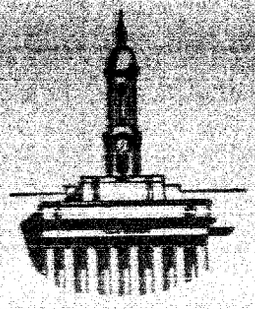
by
Thomas P. Quinn

July 1, 1964

UNPUBLISHED PRELIMINARY DATA

IONOSPHERE RESEARCH LABORATORY



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Scientific Report

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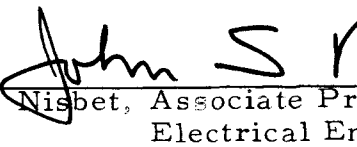
Submitted by: 

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TABLE OF CONTENTS

	Page
ABSTRACT	i
PREFACE	ii
1. INTRODUCTION	1
1.1 General Statement of the Problem	1
1.2 Previous Related Studies	2
1.2.1 Loss and Diffusion	2
1.2.2 Vertical Drifts	19
1.3 Specific Statement of the Problem	21
2. CONSTRUCTION OF PROFILES	23
2.1 Raw Data	23
2.2 Top Profile	23
2.3 Determination of the Temperature	26
2.4 Correction for Low-Lying Ionization	28
3. METHOD OF ANALYSIS	30
3.1 Calculation of the Recombination Coefficient	30
3.2 Calculation of the Diffusion Coefficient	35
4. EFFECTS OF ASSUMPTIONS USED IN THE ANALYSIS	42
4.1 The Effect of Shape Assumed for the Top Profile	42
4.2 The Effect of Determining Temperature from the Profiles	43
4.3 The Effect of the Quadratic Loss Coefficient	49
4.4 Comparison of β_{300} for N_2 and O_2 Molecular Constituents	52
4.5 Low-Lying Ionization	54
4.6 Production Due to Protonosphere Coupling	54
4.7 The Effect of a Change in Te/Ei	56
4.8 β_{300} for Individual Days	58
5. RESULTS OF THE ANALYSIS	62
5.1 Best Estimate for β_{300}	62
5.2 Best Estimate for the Diffusion Coefficient	62
5.3 Seasonal Effects	72
5.4 Drift Velocity	77

6. SUMMARY AND CONCLUSIONS 81

6.1 Assumptions in the Analysis 81
6.1.1 Top Profile 81
6.1.2 Quadratic Loss Coefficient 81
6.1.3 Different Assumptions about the Molecular Constituent 82
6.1.4 Low-Lying Ionization 82
6.1.5 Production Due to Protonosphere Content Change 83
6.1.6 Decrease in Te/Ti During the Night 83

6.2 Recombination Coefficient 84
6.2.1 The Linear Recombination Coefficient at 300 Kilometers 84
6.2.2 Seasonal Effects 85

6.3 The Diffusion Coefficient 85

6.4 Drift Velocities 86

6.5 Suggestions for Further Research 86

ACKNOWLEDGEMENTS 86a

BIBLIOGRAPHY 87

ABSTRACT

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A method by which the recombination coefficient for the nighttime F layer may be determined, and an extention of the method which allows a determination of the drift and diffusion velocities are presented.

The relative effects of different assumptions in the analysis on the calculated recombination coefficients are examined.

Recombination and diffusion coefficients are determined using data from six ionosphere stations. A strong seasonal dependence for the recombination coefficient is noted and expressions for summer and winter values are given. It is suggested that the recombination coefficient exhibits no latitude dependence.

Vertical drift velocities are found to be of the same order of magnitude as the diffusion velocities.

Author

PREFACE

In Section 1, the objective of the present work is stated and previous related works are reviewed.

In Section 2, the raw data to be used in the analysis is described along with methods for adding corrections. Corrections for effects which were neglected in reducing the original data are described.

In Section 3, methods to be used in the analysis are developed.

In Section 4, the effects of different assumptions in the analysis on the calculated value of the recombination coefficient are examined.

In Section 5, the results of the present work are presented, mostly in graphical form to facilitate their presentation, interpretation and use in future work.

In Section 6, the results are summarized and "best-estimates" for the recombination and diffusion coefficients are given. A seasonal effect exhibited by the calculated recombination coefficient is discussed.

The bibliography lists all the references mentioned in the text according to the alphabetical order of the authors.

1. INTRODUCTION

1.1 General Statement of the Problem

The ionization density distribution in the F region of the nighttime ionosphere is controlled by recombination, diffusion, drift motions of the ions and atmospheric winds.

It is important for a complete understanding of the processes involved and their diurnal, seasonal and solar cycle behavior, to be able to estimate the recombination and diffusion coefficients and the magnitude of the drift motions.

Above 200 kilometers, atomic oxygen is the major ionized constituent and it is generally accepted that the loss process is one of charge exchange with molecular nitrogen and molecular oxygen followed by dissociative recombination. The loss rate for atomic oxygen ions is thus dependent on the molecular densities N_2 and O_2, while the diffusion coefficient is proportional to the neutral density determined mainly by atomic oxygen. The result is that the predominant process in the nighttime F region is downward diffusion followed by recombination below the ion density maximum. Because of the vertical motion of the ions, recombination in the layer as a whole rather than at specific heights must be considered.

It is the purpose of this present study to use measurements from bottomside ionospheric sounders along with the best estimates available for the neutral atmospheric density distributions

and temperature to establish the processes effective in the nighttime ionosphere and make estimates of the numerical values of the important coefficients.

1.2 Previous Related Studies

1.2.1 Loss and Diffusion

A linear recombination process for F region ionization was first postulated by Bradbury (1938) to account for the upward increase in ionization above the region of maximum production. Bates and Massey (1947) suggested a double loss process of charge exchange proceeding at a rate linear in the electron density followed by dissociative recombination proceeding at a rate proportional to the square of the electron density. The complete process is then represented by the equations:

$$O^+ + X\,Y \rightarrow O\,X^+ + Y$$

$$O\,X^+ + e \rightarrow O + X$$

Ratcliffe (1956) showed that this dual mechanism would result in a loss term of the form:

$$L = \frac{\alpha\, B\, n^2}{\alpha n + \beta} \quad cm^{-3}\ sec^{-1}$$

where

α = the quadratic loss coefficient, $cm^3\ sec^{-1}$

β = the linear loss coefficient, sec^{-1}

n = electron density, cm^{-3}

At low altitudes where n is small, the loss rate is proportional to n^2, while at higher altitudes where n is larger the loss rate is linear in n.

Savitt (1950) using ionosonde data obtained at Bocayauva, Brazil before during and after the eclipse of May 20, 1947 examined two hypothesis for electron loss,

$$\text{either} \quad \frac{\partial n_e}{\partial t} = f_x q - n_e$$

$$\text{or} \quad \frac{\partial n_e}{\partial t} = f_x q - n_e^2 ,$$

where $f_x q$ is a production term for the uneclipsed fraction of the solar disk.

Using four fixed heights, 220 km., 260 km, 300 km he produced curves of electron density versus time from the ionograms for a period embracing the time of the eclipse. Ionograms were taken every 15 seconds during the eclipse and every 15 minutes for several days before and after the eclipse. By using control data before and after the eclipse he estimated

the production function.

He then obtained electron density changes with time from the n(t) curves plotted from the ionograms and also calculated $\partial n/\partial t$ using the two relations above. He concluded that the first relationship gave much better agreement with the experimental data than did the second. He found that β followed an almost linear height variation from a value of 6.5×10^{-4} sec^{-1} at 220 km to 2.8×10^{-4} sec^{-1} at 320 km. His curve indicates a value of 3.5×10^{-4} sec^{-1} at 300 km.

Ratcliffe, Schmerling, Setty, and Thomas (1956) used electron density profiles determined from ionograms taken at Slough and Watheroo for years of high sunspot number to calculate the recombination coefficient. Only mean, quiet F layer profiles were used.

Two methods were used to determine the recombination coefficient. Although they recognized that diffusion and drift could be important the methods of analyses were chosen to attempt to minimize these effects.

In the first method, n(t) curves deduced from the electron density profiles were used to determine the change of n with time at a series of fixed heights. Although the results agreed with the assumption that the loss coefficient at each height re-

mained constant for 8 hours, the accuracy was not sufficient to decide with certainty whether the loss rate was proportional to n or to n^2. However, selected nights during which the electron density decreased smoothly at all heights were used to determine a linear loss rate. Results from a total of 40 nights for Slough and 3 nights for Frosenburg were presented. The data for heights between 250 km and 350 km were shown to be fitted fairly well by the relation,

$$\beta = 10^{-4} \exp\left(\frac{300-h}{50}\right) \text{ sec}^{-1} \quad .$$

In the second method the continuity equation

$$\frac{\partial n}{\partial t} = q - \beta f(h) n^a - \frac{d(nV)}{dh}$$

was integrated. It was assumed that if the "mean quiet F layer" is repeated from one day to the next, then the total production in a unit column of atmosphere for a 24 hour period would be equal to the total loss, quite independent of any motions. However, since electron density profiles extend only to the peak of the layer the integration could not be carried out above this level. It was therefore assumed that in a 24 hour period the movement of ionization across this upper boundary would average to zero. A Bradbury hypothesis (in which production is assumed due to a single constituent with a single maximum of production) and a Chapman hypothesis (in which

a double peak of production is assumed) were both used in the analysis. By choosing $a = 1$ a linear loss law was assumed and by choosing $a = 2$ a quadratic loss law was assumed. The ratio of total production to total loss in the vertical column for 24 hours was then calculated for different times, seasons and places. The joint hypotheses which gave the most consistent results for this ratio were chosen to be most likely correct. It was concluded that the Bradbury hypotheses for production and the previously stated loss law were most likely to be correct.

Having accepted the Bradbury production hypothesis and using the deduced relation for the loss rate, the continuity equation was rewritten as

$$\frac{\partial n}{\partial t} = q - \beta n - M.$$

By calculating n and $\frac{\partial n}{\partial t}$ from the experimental results, β from the deduced relation, and q from the production function, it was then possible to compute the flux term M. Estimates of 100 cm^{-3} sec^{-1} near midday and 40 cm^{-3} sec^{-1} near midnight for M were given. By using the relationship given by Ferraro (1945) and a parabolic fit to the electron density profiles it was then possible to estimate the neutral atmospheric density. The resulting values for the neutral density were shown to be a factor of 5 higher than those suggested by Bates (1954) from rocket results.

(It should be pointed out that the neutral densities estimated in this paper are close to present day estimates of the neutral density.

The early rocket results have since been shown to be too low.)

It was also suggested that when the peak of the F_2 layer is high, its level is determined by the diffusion process.

Van Zandt, Norton and Stonehocker (1960) calculated a recombination coefficient based on data obtained by ionosonde at Danger Islands in the central Pacific during the eclipse of October 12, 1958. Soundings were made every 15 minutes between 0000 and 0530 hours, every one minute between 0530 and 1205 and again every fifteen minutes between 1205 and 2400 on the day of the eclipse. First contact of the eclipse at 300 km occurred at 0734 hours local time. Soundings were also made for 6 control days before and 5 control days after the eclipse. The day of the eclipse was quiet in both solar and geomagnetic activity.

Curves of n(t) were plotted from the reduced electron density profiles at fixed heights at 10 km intervals. To analyze these data it was assumed that changes in electron density during the eclipse were due entirely to production of electrons by photoionization and loss by recombination. Photo ionization and loss using only the constituents 0 and N_2 respectively were considered. A linear loss process was used.

With the above assumptions, the continuity equation was written in the form

$$\left[\frac{1}{n}\ \frac{\partial n}{\partial t}\right] = q\left[\frac{E}{n}\right] - \beta$$

where

E = uneclipsed fraction of the sun's disk

q = production function

n = ion density

β = recombination coefficient

In the above equation n and E are functions of both height and time while q and β were assumed to be functions of height only.

The bracketed terms were determined from the eclipse data and plotted against one another, the slope of the plot thus yielding a value for q and the intercept a value for β. From the results the following expression for β was deduced.

$$\beta(h) = 6.8 \times 10^{-4} \exp\left(-\frac{h - 300}{103}\right) \sec^{-1}.$$

The authors noted that this value was higher than previous estimates and would be unreasonable for the nighttime F region. They therefore concluded that either the rate of recombination varies diurnally or else the nighttime rate is much larger than the apparent rate determined by previous workers.

Martyn (1956) investigated the possible effects of diffusion processes at F region heights by investigating the continuity equation in the absence of production. He assumed that the ion and electron densities were equal and that transport is along the earth's magnetic lines of force.

He then showed that the velocity due to gravity and pressure gradient is given by:

$$v = - \frac{g}{\nu_i} \left[1 + \frac{2}{n} \frac{\partial n}{\partial z} \right] \sin^2 I$$

g = accel. of gravity

ν_i = ion neutral collision frequency

n = ion density

z = height measured in scale heights

I = magnetic dip angle

This result was also determined by other workers but by more complex reasoning.

He concluded that in the absence of production and loss, for an ionized layer to move vertically or remain stationary without changing shape, the velocity must have no height gradient. Since the Chapman C_α function given by

$$n = n_m \exp \frac{1}{2} (1 - z - \exp - z)$$

satisfied the relevant differential equation, it was concluded that it must be unaffected by diffusion.

The more practical case of the F region at night controlled by loss and diffusion was also investigated. However, it was assumed in this analysis that the recombination coefficient and the diffusion coefficient had the same height dependence which would only be true if the molecular constituent involved in the loss process of charge exchange had the same scale height as atomic oxygen.

The results of this investigation lead to the following conclusions:

1. Any arbitrary ionization distribution will redistribute itself to form a C_α region in time. The time required at F_2 heights is a few hours if the pre-rocket density estimates are correct and only a few minutes if rocket measured densities are correct.

2. The observed behavior of the F_2 region is controlled by a linear loss process which decreases with height and by diffusion.

3. Rocket measurements of atmospheric densities at heights above 150 km appear to be subject to a gross systematic error and are much lower than can be reconciled with other types of observation.

4. If the decay of ionization follows a linear law than a stable C_α region will tend to form at 360 km if pre-rocket densities are accepted which is observed to occur at night throughout the world except at the geomagnetic equator.

5. The formation of a C_α region by diffusion a few hours after sunset accounts for the well known fact that the under surface of the F_2 region assumes a parabolic height distribution.

Dungey (1956) examined the effect of ambipolar diffusion of electrons and ions on the behavior of the F layer at night. He made the following assumptions:

(1) Production is zero,

(2) the region is isothermal,

(3) the proportion of different neutral constituents is independent of height; the scale height is H and $z = {}^{h}/H$,

(4) the positive ions have the same composition as the neutral constituents,

(5) The downward diffusion velocity is inversely proportional to the neutral density and is given by

$$v = be^{z}\left(1 + \frac{2}{n}\,\frac{\partial n}{\partial z}\right)$$

where b is constant.

(6) the loss rate is given by

$$L = \beta e^{-kz}$$

where k and β are constants,

(7) horizontal variations are neglected

(8) the velocity, V, of charged particles due to causes other than diffusion is vertical and uniform.

Under these assumptions he examined the continuity equation including loss, diffusion and drift terms.

It is first shown that for layer-shape preserving solutions, k must be positive. For $k = 0$, and $V < 0$, the boundary conditions are not satisfied.

For $k = 0$, and $V = 0$, Yonezawa (1955), quoted by Dungey, found that the limiting solution was a Chapman layer which falls but does not change shape.

For $k = 0$, and $V > 0$, the solution is a Chapman layer which does not fall but whose height is determined by $be^{z_{max}} = V$.

For the special case of $k = 1$, the shape is a Chapman layer, and the loss of ionization tends to move the layer upwards.

The case of $k = 2$ was studied in detail because of the grounds for believing that this approximates to reality. For this case, the profiles obtained by imposing the shape preserving restriction of

$$\frac{\partial n}{\partial t} = - \lambda n$$

where λ is constant,

were found to have the same general form as the Chapman layer.

n varied as $e^{-z/2}$ at great heights, had a maximum lower down, and became very small about 2 scale heights below the maximum. The part of the profile below n_{max} closely resembled a parabola down to $^1/3\ n_{max}$.

Using electron density profiles for Slough, England, and taking an overall median value of 0.15 per hour for λ, it was concluded that β_{max} was approximately 0.22×10^{-4} sec^{-1} at 370 km. The corresponding scale height for atomic oxygen was 70 km.

For the above values and assuming O_2 to be the molecular constituent the value of β at 300 km would be 1.6×10^{-4} sec^{-1}.

He pointed out that although the layer profiles calculated for $k = 2$ seem to closely approximate parabolas below n_{max}, the parabolic approximation is inadequate and can lead to large errors when attempts are made to relate the recombination and diffusion coefficients to data obtained from the parabolic approximation. This indicates the importance of obtaining accurate electron density profiles from ionogram data.

A study of the dependence of $\frac{d}{dt}$ (log n_{max}) on h_{max} was also made and graphically illustrated for low, medium and high sunspot numbers. For medium sunspot numbers this data could be fitted by an exponential curve with a scale height of approximately 0.7 H.

Dalgarno (1958a, 1961) suggested that the diffusion coefficient deduced by Ferraro (1945) and used by many workers may be too high by a factor of 4 for application to the F region. He based this conclusion on the fact that the expression used by Ferraro for the temperature dependence of the ion diffusion cross section was based on a mobility measurement for N_2^+ diffusing in N_2. In addition to the point that in the F region the diffusion depends primarily on the diffusion coefficient for O^+ in O, the ions diffusing in the measurement were almost certainly N_4^+.

Dalgarno's values were based on an empirical method which was derived from previous mobility studies, Dalgarno (1958b). In this mobility study he points out that for an ion diffusing in its parent gas the diffusion cross section at low temperatures is approximately twice that for the charge transfer process. At higher temperatures the diffusion cross section is approximately

$$Q_d = 2(Q_t + Q_p)$$

where Q_p is a cross section dependent on impact parameters which was calculated and plotted by Dalgarno, McDowell and Williams (1958). However since no experimental information for Q_t or the mobility is available it is not possible to determine the diffusion coefficient for O^+ in O directly. By assuming a mobility of 1.6 cm^2 $volt^{-1}$ sec^{-1} at 300°K the following normalized values for the diffusion coefficient and collision frequency at 300°K were determined.

(2) at the peak, and below it, the electron density is given quite accurately by the expression $N = q/\beta$ just as though there were no diffusion;

(3) at levels more than one scale height above the peak, the distribution is controlled by diffusion and takes an exponential form.

(4) near the peak the distribution is approximately parabolic with thickness inversely proportional to the gradient of of the function $\log \left(\frac{\beta H^2}{D} \right)$.

(5) The effect produced by a vertical drift depends on the ratio of its magnitude to D/H. An upward drift for which this ratio is unity raises the peak by about one scale height and increases the peak electron density; a downward drift has opposite effects. The shape of the layer is not greatly altered.

These conclusions are not materially effected if the scale height of the atmosphere varies with height.

By making use of the relation $\frac{H^2 \beta_{max}}{D_{max}}$ which specifies the height of the layer maximum, as suggested by Rishbeth and Barron, then if the height of the layer and H^2 are known, either β_m or D_m may be found if one or the other is known or assumed. This was done by Bowhill (1961).

Rishbeth (1963) used an analog computer to simulate the behavior of the F region under various conditions. This model study confirmed that a slab of ionization introduced into the ionosphere eventually forms a stationary distribution whose shape is independent of the height at which the ionization is introduced. He concluded that, as is the case for equilibrium layers, for the nighttime F region the peak of a stationary distribution lies at a height where the ratio $\frac{H^{2\beta}_{max}}{D_{max}}$ lies between 0.15 and 0.3 for the models he considered.

Nisbet and Quinn (1963) determined the recombination coefficient for the nighttime F region by considering the decay of ionization within a tube, bounded by lines of the earth's magnetic field and extending from the bottom of the layer to the geomagnetic equator. It was shown that by considering the recombination in the entire layer in this way the flux terms in the integrated continuity equation due to motions could be neglected.

Monthly mean electron density profiles for Puerto Rico published by the Central Radio Propagation Laboratory of the National Bureau of Standards were corrected for low-lying ionization and used to determine the bottomside of the layer. The portion of the layer situated above the peak was represented by a Chapman layer to 500 km., and from 500 km to the geomagnetic equator by

an equilibrium distribution for 0^+, He^+, and H^+ ions

It was pointed out that of the total ionization recombining below the peak of the layer, sixty percent of the ions have diffused down from above the peak.

The importance of diffusion and transport in the analysis of nighttime recombination parameters was thus made clear

It was found that the recombination coefficient is temperature dependent and varies as much as 30:1 over a solar cycle as does the 0_2 density Values for summer months were noted to be consistently higher than those for winter months

Expressions for the recombination coefficient for three temperatures were given:

$$\beta = 1.5 \pm 1 \times 10^{-5} \exp\left(-\frac{Z-300}{20}\right) \sec^{-1} \text{ for } 700^{\circ}K$$

$$\beta = 6.5 \pm 2 \times 10^{-5} \exp\left(-\frac{Z-300}{29}\right) \sec^{-1} \text{ for } 1000^{\circ}K$$

$$\beta = 19 \pm 6 \times 10^{-5} \exp\left(-\frac{Z-300}{37.5}\right) \sec^{-1} \text{ for } 1300^{\circ}K$$

For the case of β proportional to the molecular oxygen density:

$$\beta = \gamma n(0_2),$$

an estimate of $\gamma = 4.5 \pm 1.5 \times 10^{-12}$ cm^3 sec^{-1} was also presented.

Nisbet (1963) used the results of Dungey (1956) to study the nighttime F layer by examining $\frac{\partial n_{max}}{\partial t}$. After modifications were applied to take into account the change in layer thickness and in neutral atmospheric temperature it was shown that the equilibrium layer theory gave values of recombination coefficient which were in substantial agreement with the results obtained by Nisbet and Quinn (1963).

1.2.2 Vertical Drifts

The concept of a vertical drift motion of ionization caused by an electrodynamic force was introduced by Martyn (1947 a,b). According to this "dynamo" theory a current flows in the lower ionosphere due to tidal motions, Baker and Martyn (1953). The electrostatic field is coupled to the F region by means of the high conductivity along the lines of the earth's magnetic field. This electrostatic field and the perpendicular magnetic field provide a driving force for drift of ions and electrons in the F region.

Drift motions of ionization have been studied using this and related theories by Martyn (1953, 1955), Maeda, K. (1952, 1953), Fejer (1953), Maeda, H. (1955) and others. Vertical drift velocities calculated by these workers were usually between 10 and 20 meters/sec.

Reviews of the effects of the tidal and electromagnetically induced drifts have been given by Martyn (1959) and by Hines (1959).

Ionization drift velocities based on the dynamo theory are usually calculated by relating variations in the earth's magnetic

field as measured at the earth's surface to currents flowing in the lower ionosphere. Assumptions about tidal motions and estimates of the conductivity are clearly necessary in such calculations assuming the theory itself is correct. Some workers have therefore developed methods to calculate the drift velocity independent of assumptions about the driving force causing the motion.

Chandra, Gibbons and Schmerling (1960) calculated the vertical velocity by integrating the continuity equation from a reference height up to the height at which the velocity is to be found. The production and loss terms due to Ratcliffe et. al. (1956) were used and an arbitrary velocity at the starting height was assumed. By adjusting these parameters it was possible to get velocity profiles which agreed with those predicted by the dynamo theory. No attempt was made to separate the effects of diffusion from the calculated velocity and it was shown that by changing the production by a factor of 2 the resulting velocity would change by ± 100 %.

Garriott and Thomas (1962) calculated vertical drift velocities by expanding the movement term in the continuity equation and then

calculating the drift velocity required to balance the equation at each height. Data taken from ionograms were smoothed and a numerical method used to evaluate both the time and space derivatives appearing in the equation using these data. A constant positive scale height gradient, a neutral atmosphere model, and Ferraro's expression for the diffusion coefficient were used throughout the study. Although their solutions tend to minus infinity near the bottom of the layer and to plus infinity about a scale height higher, numerical values were obtained for the region between these limits. Their values were of the order of tens of meters/second, and for disturbed days reached as high as 80 meters/second.

1.3 Specific Statement of the Problem

It is clear from the preceding summary of F region studies that the diffusion coefficient is known at best to within an order of magnitude. Recent studies have provided a better estimate of the recombination coefficient, but the way in which it varies with other parameters and its seasonal behavior need further study. Much less is known about the drift velocity in the nighttime F region primarily due to the difficulty of separating its effects from those due to diffusion.

In the present study a method which does not depend on neglecting either drift or diffusion will be used to study the variations in the recombination coefficient, and to determine the drift

and diffusion velocities from which the diffusion coefficient can be estimated.

2. CONSTRUCTION OF PROFILES

2.1 Raw Data

The calculation of the recombination coefficient and other ionospheric parameters requires a knowledge of the height dependence of the ionization density. In this study monthly mean electron density profiles calculated by the Central Radio Propagation Laboratory of the National Bureau of Standards are used as the basic measured data. To construct these profiles daily electron density profiles are first reduced from ionosonde data by the Budden matrix method. These electron density profiles are then extended beyond the layer maximum by adding a Chapman layer with a constant scale height of 100 km . At the lowest heights a simple extrapolation is used. The daily profiles are then averaged for days when the magnetic K_p index is less than 4.5.

Averaged profiles for the period March, 1959 through February, 1962 for the stations: Puerto Rico; Grand Bahama Island; White Sands, New Mexico; Fort Mommouth, New Jersey; Adak, Alaska; and St. John, Newfoundland were provided by N.B.S. for this study.

These profiles were only used to represent the layer up to the electron density maximum. Beyond this height a top profile was added as described in the next section.

2.2 Top Profile

The electron density profiles described in Section 2.1 can be used to represent the ionization density up to the height of the layer maximum. But to calculate the recombination coefficient a

knowledge of the layer extending to the geomagnetic equator, G.E., is required. It is therefore necessary to add a "top" profile which describes the ionization density distirbution between the layer maximum and the G.E.

Dungey (1956) and Rishbeth and Barron (1960) have shown that a Chapman C_α function should provide a good fit to the electron density profile near the peak of the layer. The region between the layer peak and 500km was therefore represented by a Chapman C_α function.

Above 500 km helium and hydrogen ions become important and their influence on the atomic oxygen ion distribution must be considered. Mange (1960) has shown that under equilibrium conditions the distribution of density of an ion of mass m_i at geopential altitude Z is given by

$$n_i = n_{io} \exp - \int_{Z_o}^{Z} \left(\frac{m_i - \frac{m_+}{2}}{kT} g \right) dZ \qquad (2.2.1)$$

where m_+ = the mean mass of the positive ions

m_i = mass of i th ion

n_{io} = density of ion i at $Z = Z_o$

g = acceleration due to gravity

k = Boltzmann's constant

T = temperature, ^oK

Z = geopotential height

Bauer (1961) used this result to develop an expression for the electron density in the presence of two species of ion, and Nisbet, Quinn, and Carson (1961) generalized the result to include three ionic species which more closely approximates reality. Using the result of Nisbet et. al., the atomic oxygen ion density can be written:

$$n = \frac{n_o \sqrt{1 + \lambda + \psi}\ \exp\left(-\frac{Z - Z_o}{H(0^+)}\right)}{\sqrt{\exp\left(-\frac{Z-Z_o}{H(0^+)}\right) + \lambda\ \exp\left(-\frac{Z-Z_o}{H(He^+)}\right) + \psi \exp\left(-\frac{Z-Z_o}{H(H^+)}\right)}} \qquad (2.2.2)$$

where

$$\lambda = \frac{n_o(H_e^+)}{n_o(0^+)} \qquad (2.2.3)$$

$$\psi = \frac{n_o(H^+)}{n_o(0^+)} \qquad (2.2.4)$$

The quantities λ and ψ are temperature dependent, but an estimate of their variation with temperature can be made by using the curves due to Bauer (1962) which show the temperature dependence of the heights at which the ratios $n(H^+e)/n(0^+)$ and $n(H^+)/n(H_e^+)$ are unity.

A polynomial fit to these curves was made and the quantities ψ and λ calculated in a computer program for the height of 500 km. The ion density given by equation 2.2.2 was then used to represent

the distribution for atomic oxygen ions above 500 km .

2.3 Determination of the Temperature

Because the distribution of both the neutral atmospheric density and the atomic oxygen ion density depend upon the temperature, a means for determining the temperature must be used.

Jacchia (1961) has shown that the temperature above the thermopause as determined from satellite retardation measurements is well correlated with measurements of the 10.7 cm solar flux, the nighttime minimum temperature being given by:

$$T_{Min} = \left(555^{o} + 3S\right)\ ^{o}K \qquad (2.3.1)$$

where S is the monthly mean of the 10.7 cm solar flux in units of 10^{-22} watts/m^2/cycle/sec. The diurnal temperature variation was found to be

$$T_{day} = 1.33\ T_{night} \qquad (2.3.2)$$

Harris and Priester (1962) adopted the following relation for the nighttime temperature,

$$T_{Min} = \left(275 + 4.5\ S\right)\ ^{o}K \qquad (2.3.3)$$

Their diurnal change of 1.58 was slightly larger than that suggested by Jacchia.

Jacchia (1962) introduced a correction for the "27 day" oscillation of E.U.V. heating and for the solar wind. Including

these effects, the temperature is given by:

$$T_{night} = 330 + 2.1\,S + (2.0 + 0.5 \cos 4\pi \frac{\tau - \text{Apr.}7}{365})\,\bar{S}\ ^{o}K \qquad (2.3.4)$$

$$T = T_{night}\,(1 + 0.4 \cos^{4} \frac{\phi}{2}) + 1.2\,A_{p}\ ^{o}K \qquad (2.3.5)$$

where

S = daily value of 10.7 cm solar flux

$\bar{S}$ = monthly mean of 10.7 cm solar flux

τ = time in days

ϕ = geocentric angular distance from the center of the diurnal bulge

A_p = 3 hour geomagnetic index

Nicolet (1963) examined the correlation between the atmospheric temperature and both the 10.7 cm and 8 cm solar fluxes. He concluded that for the 10.7 cm solar flux the nighttime temperature is given by:

$$T_{night} = 380^{o} + (4 \pm 0.25)\,S\ ^{o}K \qquad (2.3.6)$$

for $S < 200$.

He showed that the "27 day" oscillation must be taken into account in calculating daily temperatures from the solar flux.

For S numbers greater than 100, the difference between the

temperature derived using the Harris and Priester model and that from the Nicolet model is less than 10 percent. However a diurnal temperature variation is also given by Harris and Priester, and in order to include a diurnal temperature variation in this study, the temperature is based on the Harris and Priester formula. In one case, to examine daily values of β_{300} in Section 4.8, equations (2.3.4) and (2.3.5) due to Jacchia are used.

2.4 Correction for Low-Lying Ionization

The electron density profiles described in Section 2.1 are in error at the lower heights due to low-lying ionization which was not taken into account when the ionograms used were reduced. A correction suggested by Long (1962) was therefore applied to the raw data before it was used in the analysis. The following table presented by Long was used to determine the amount by which each height should be reduced for a given electron density. For values not shown in the table an interpolation by curve fitting was used.

Table 1

Correction for Low-lying Ionization
(After Long (1962))

The amount in kilometers by which the height of the given electron density n (calculated assuming no ionization with plasma frequency < f min) should be lowered to make an approximate allowance for the effect of the low-lying ionization in the upper ionosphere n-h curve.

Month	Latitude	Error in kilometers, due to the neglect of low lying ionization, in the calculated height of the following electron densities $n \times 10^{-5}$ cm^{-3}:	0	1	2	3	4	5
Dec. 1957	>30°		120	21	15	12	8	5
	<30°		30	16	11	8	7	5
June 1958	>30°		100	16	10	9	8	5
	<30°		20	11	5	4	3	3

3. METHOD OF ANALYSIS

3.1 Calculation of the Recombination Coefficient

The method used to calculate the recombination coefficient is essentially that presented by Nisbet and Quinn (1963). The details are presented here and the method is extended in Section 3.2 to allow calculation of the diffusion coefficient.

In the absence of production, the continuity equation for the F region may be written

$$\frac{\partial n}{\partial t} = - L - \text{div}\,(n V) \qquad (3.1.1)$$

where

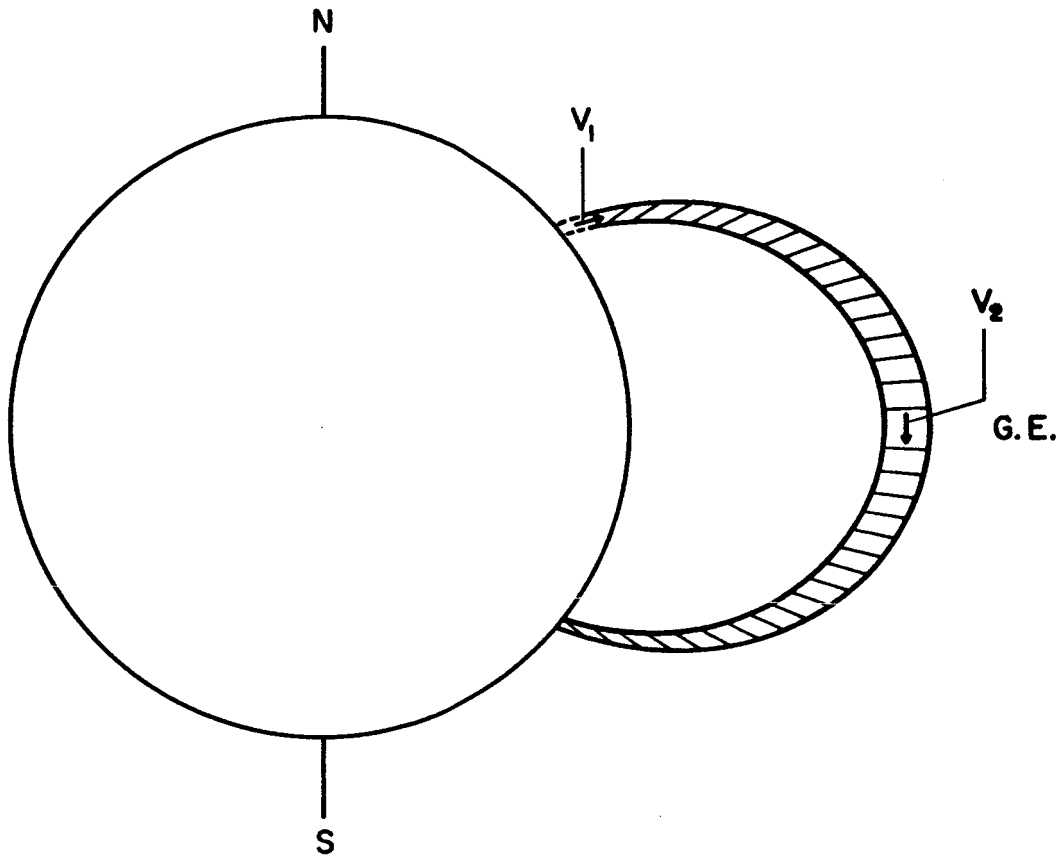
n = electron or ion density

L = rate of loss of ionization

V = transport velocity

Equation (3.1.1) may be integrated along a tube of volume bounded by lines of the earth's magnetic field and enclosing unit area at a height of 300 km. The method is illustrated in Figure 1. If motion of the ions is restricted to be along field lines, only those ions passing through either end of the tube can contribute to the divergence term. Integrating equation (3.1.1) then gives the expression:

$$\frac{\partial}{\partial t} \int_{Z_1}^{Z_2} n \, dv = - \int_{Z_1}^{Z_2} L \, dv + n_1 a_1 V_1 - n_2 a_2 V_2 \qquad (3.1.2)$$



$$\beta_{300} = \frac{\int_0^{G.E.} n\,dV\Big|_{t_1} - \int_0^{G.E.} n\,dV\Big|_{t_2}}{\int_{t_1}^{t_2}\int_0^{G.E.} ne^{-\frac{Z-300}{H(xy)}}\,dV\,dt}$$

METHOD OF CALCULATING β_{300} BY RESTRICTING IONIZATION TO A TUBE OF FORCE

FIGURE 1

where n_j, a_j, V_j, are the electron or ion densities, cross sectional area of the tube, and ionization velocity respectively at the height Z_j, and dv is the differential volume element.

Choosing the lower limit of the layer at an altitude such that n is sufficiently small allows the contribution of the term $n_1 a_1 V_1$ to be disregarded. Taking the upper limit to be at the geomagnetic equator, G.E., the net transport of ions across this boundary is zero by symmetry between the northern and southern hemispheres.

The remaining terms of equation (3.1.2) may now be integrated with respect to time.

$$\int_0^{G.E.} n\,dv \Big|_{t_1} - \int_0^{G.E.} n\,dv \Big|_{t_2} = \int_{t_1}^{t_2} \int_0^{G.E.} L\,dv\,dt \qquad (3.1.3)$$

Thus if the electron or ion density as a function of altitude and time, and the height dependence of L are known, L may be calculated at any height.

Considering the dual loss mechanism of charge exchange followed by dissociative recombination, the loss rate may be written:

$$L = \frac{\alpha B\, n^2}{\alpha n + \beta} \qquad (3.1.4)$$

where

β = rate coefficient for linear loss,

α = rate coefficient for quadratic loss

Because the under surface of the F region at night has a large vertical gradient of electron density with height, the total number of electrons recombining in the region where the loss rate αn^2 is dominant is a small fraction of the total. As a first approximation, the loss rate is therefore given by:

$$L = \beta n. \tag{3.1.5}$$

An iterative technique was used to modify the loss term to the form of equation (3.1.4) after an approximate value of β had been determined.

Above a height of 200 km, the F region is practically isothermal at night, Nicolet (1960). Also, several hours after sunset the molecular oxygen and molecular nitrogen densities to which β may be assumed proportional correspond to diffusive equilibrium profiles. The height dependence for β may therefore be written:

$$\beta = \beta_{300} \exp - \left(\frac{Z-300}{H(xy)} \right) \tag{3.1.6}$$

where

β_{300} = numerical value of β at 300 km.

H(xy) = scale height of either molecular nitrogen or molecular oxygen.

In the present analysis H(xy) has been taken to be that for molecular oxygen, except in Section (4.4) where a comparison of the results for the two different densities is made. Combining

equations (3.1.3), (3.1.5), and (3.1.6) the recombination coefficient at 300 km is given by:

$$\beta_{300} = \frac{\int_o^{G.E.} n\, dv \Big|_{t_1} - \int_o^{G.E.} n\, dv \Big|_{t_2}}{\int_{t_1}^{t_2} \int_o^{G.E.} n \exp - \left(\frac{Z-300}{H(xy)}\right) dv\, dt} \qquad (3.1.7)$$

Using the profiles for the atomic oxygen ion density as a function of height for successive hours throughout the night, the volume integrals in equation (3.1.7) may be computed. If the successive hours are labeled 1 to m, then β_{300} may be evaluated:

$$\beta_{300} = \frac{\int_o^{G.E.} n\, dv \Big|_{t_1} - \int_o^{G.E.} n\, dv \Big|_{t=m}}{3600 \sum_{t=\frac{1}{2}}^{t=m-\frac{1}{2}} \int_o^{G.E.} n \exp - \frac{Z-300}{H(xy)}\, dv} \qquad (3.1.8)$$

The resulting value for β_{300} has been computed neglecting the αn^2 loss rate term in equation (3.1.4) which is equivalent to taking $\alpha = \infty$. This value can now be used to provide a more accurate estimate of the total loss rate at lower altitudes by forming a new loss rate term:

$$\frac{L}{\beta_{300}} = \frac{\alpha n^2 \exp - \left(\frac{Z-300}{H(xy)}\right)}{\frac{\alpha n}{(\beta_{300})_{\alpha=\infty}} + \exp \left(\frac{Z-300}{H(xy)}\right)} \qquad (3.1.9)$$

The calculation is then repeated using this modified loss term for 3 iterations. It has been found that the iteration converges rapidly with 2 trials and the process was therefore repeated only 3 times.

In order to perform these calculations using the great amount of available data, the equations were programmed for computation by an I.B.M. 7074 digital computer. The electron density profiles were read into the computer, corrected for low lying ionization, a top profile added and β_{300} computed for each month's data.

3.2 Calculation of the Diffusion Coefficient

Once the recombination coefficient has been determined, it can be substituted into equation (3.1.2) and the integration carried out from the base of the layer up to the height at which the total velocity is to be found.

$$\frac{1}{t_2 - t_1}\left(\int_h^{G.E.} ndv\Big|_{t_1} - \int_h^{G.E.} ndv\Big|_{t_2}\right) = \int_h^{G.E.} L\,dv\Big|_{t_{12}} - n_h a_h V_h \qquad (3.2.1)$$

and solving for V_h

$$V_h = \frac{\frac{1}{t_2 - t_1}\left(\int_h^{G.E.} ndv \Big|_{t_1} - \int_h^{G.E.} ndv \Big|_{t_2} \right) - \int_h^{G.E} L\,dv \Big|_{t_{12}}}{n_h \quad a_h} \qquad (3.2.2)$$

The velocity V_h is the total velocity of ionization entering from the lower end of the tube including both drift and diffusion components.

Beginning at the bottom of the layer, the terms in equation (3.2.2) were computed to 10 kilometer height intervals up to the peak of the layer, thus producing a vertical velocity-height profile for the layer. Electron density profiles for hourly intervals between 2300 and 0200 hours were used. The times t_2 and t_1 were chosen two hours apart and the loss term evaluated for the intermediate hour. Four velocity-height profiles were thus computed for each night.

The velocity V_h contains both the drift and diffusion components and a method to separate the two is now necessary in order to determine the diffusion coefficient.

The diffusion velocity due to gravity and partial pressure gradient is given by:

$$V_d = - \frac{2Hg}{\nu_i} \left[\frac{1}{2H} + \frac{1}{n} \frac{\partial n}{\partial z} \right] \qquad (3.2.3)$$

where ν_i = ion neutral particle collision frequency.

In terms of the diffusion coefficient, D

$$V_d = - D \left[\frac{1}{2H} + \frac{1}{n} \frac{\partial n}{\partial z} \right] \sin^2 I \qquad (3.2.4)$$

where

$$D = \frac{2 H g}{\nu_i} \qquad (3.2.5)$$

Assuming atomic oxygen to be the major neutral constituent, equation (3.2.3) may be written:

$$V_d = \frac{-2 Hg \sin^2 I}{\nu_o} \exp \left(\frac{Z - Z_o}{H} \right) \left[\frac{1}{2H} + \frac{1}{n} \frac{\partial n}{\partial z} \right] \qquad (3.2.6)$$

and this expression may be normalized to give

$$V_{dn} = \frac{V_d \nu_o}{2Hg \sin^2 I} = - \exp \left(\frac{Z - Z_o}{H} \right) \left[\frac{1}{2H} + \frac{1}{n} \frac{\partial n}{\partial Z} \right] \qquad (3.2.7)$$

This normalized diffusion velocity may now be computed from the electron density profile data. A five point numerical differentiation formula was used to compute the derivative term $\frac{\partial n}{\partial Z}$.

A normalized diffusion velocity-height profile was calculated using equation (3.2.7) for each of the 4 hours for which a total velocity profile had been calculated.

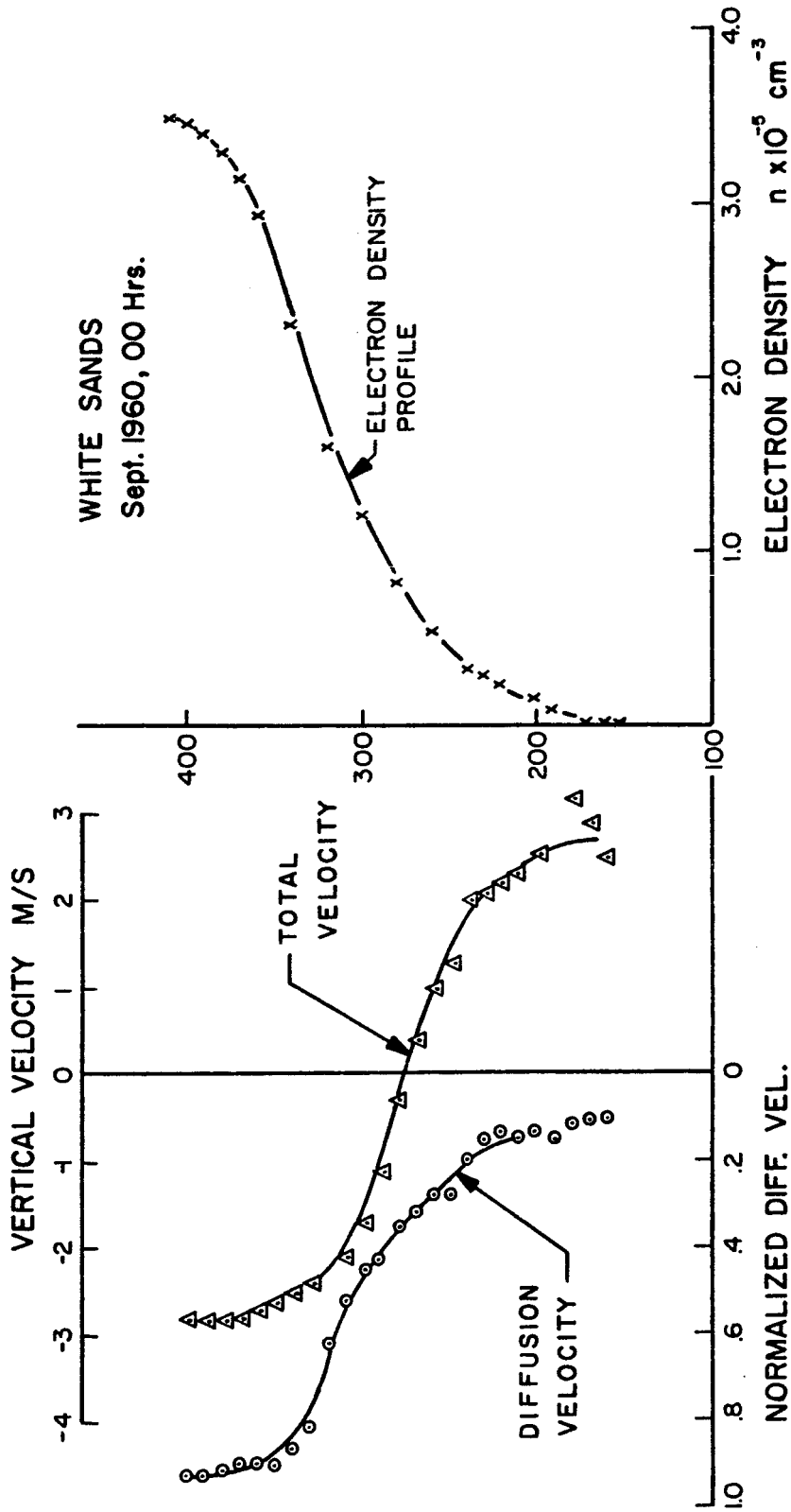
If the drift component of the total velocity is independent of height then the normalized diffusion velocity profile should match the total velocity profile.

An example of the two velocity profiles is shown in Figure 2. The scale of the normalized diffusion velocity profile is arbitrary and is chosen so that when the two curves are overlapped, the separation in the scale zero's is a measure of the constant drift velocity. In the example of Figure 2 this drift velocity would be approximately 3 meters/second. Another example is shown in Figure 3 where the drift velocity is approximately zero.

If the drift velocity determined by the above technique is subtracted from the total velocity the remaining velocity is due to diffusion and its value at the maximum of the layer is then known. The diffusion coefficient may thus be determined by substituting the diffusion velocity into equation (3.2.4).

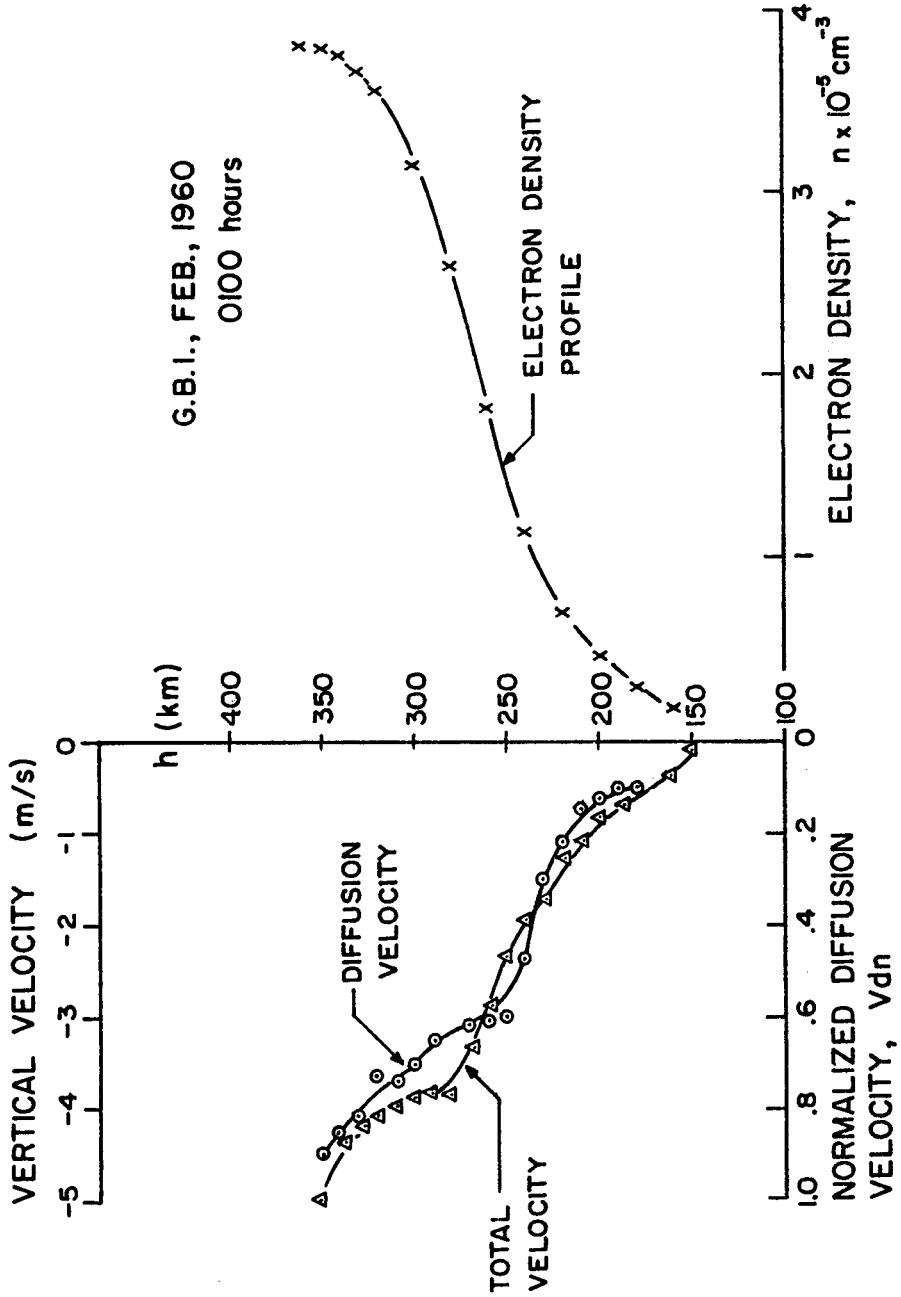
$$D_{max} = \frac{2H\ V_d}{\sin^2 I} \qquad (3.2.8)$$

Because the diffusion coefficient depends on the neutral atmospheric density it is often expressed in terms of its product with the neutral density and in this form the diffusion coefficient is given by:



DIFFUSION AND TOTAL VELOCITY PROFILES WITH Vdrift APPROXIMATELY 3 METERS/SEC

FIGURE 2



DIFFUSION AND TOTAL VELOCITY PROFILES WITH Vdrift APPROXIMATELY ZERO

FIGURE 3

$$D\, n(M) = \frac{2H\, V_d\, n(M)}{\sin^2 I} \qquad (3.2.9)$$

where n(M) = neutral atmospheric density.

4. EFFECTS OF ASSUMPTIONS USED IN THE ANALYSIS

In this section the effects of assumptions about different quantities entering into the calculation of the recombination coefficient are discussed. For these calculations selected months with calculated values of β_{300} near the mean value of β_{300} for all seasons were used except where noted differently.

4.1 The Effect of Shape Assumed for the Top Profile

The method of adding a top to the electron density profile was discussed in Section (2.2). It is of interest to examine the effects of different assumptions about this top profile on the calculated value of β_{300}. So that estimates may be made of the effects of the assumed helium and hydrogen ion distributions and the bounding of the ionization by field lines on calculated values of β_{300} a model was chosen in which the electron density followed a Chapman C_α function above the peak:

$$n = n_{max} \exp \tfrac{1}{2} \left(1 - \frac{Z - Z_m}{H} - \exp - \frac{Z - Z_m}{H} \right)$$

where Z is the geopotential height

and Z_m is the geopotential height of the maximum

H is the scale height of atomic oxygen

n_m is the maximum electron density

Values for several months which the complete analysis had shown to give close to the mean value of β_{300} were recalculated

with this model top and the results are shown in Figure 4. It is apparent that the values of β_{300} are about 20% higher for the C_α top. This is as would be expected for the presence of the helium ions causes the atomic oxygen ions to fall off more rapidly with height than they would in their absence.

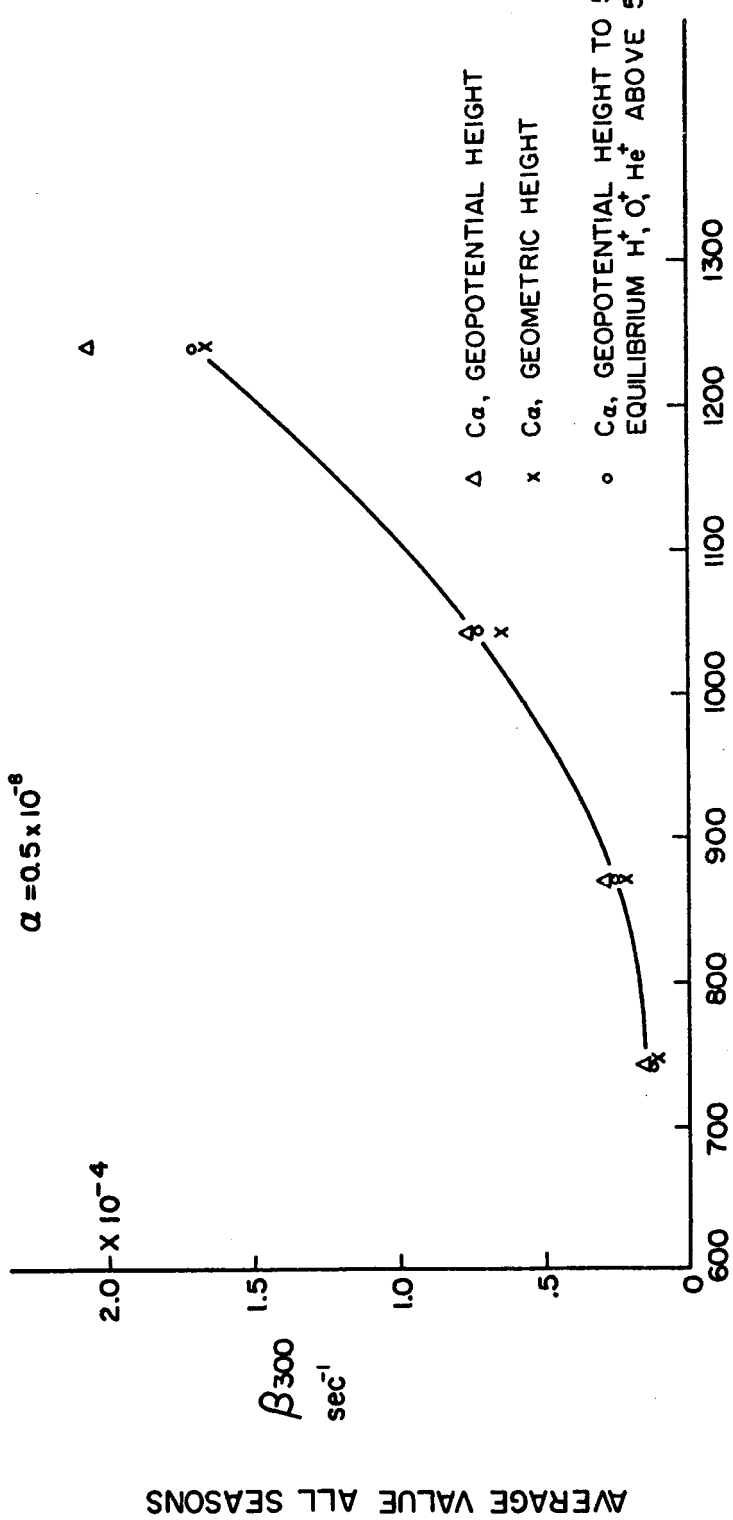
Various model studies such as those of Dungey (1956) find it convenient to use a top in which both the effects of field line trapping and decreasing g with altitude are neglected. To study the effect of these additional approximations a C_α model was considered in which the geometric height was used. The same months were used for these calculations and the results are shown in Figure 4. It is apparent that for Puerto Rico the effect of the various approximations approximately offset one another and the final results are quite close to those of the more complete analysis.

It may be concluded that provided the layer shape remains uniform and the effects of temperature changes are included, the measurements of β_{300} are not critically dependent on the exact form of the relation used for the electron density in terms of the height.

4.2 The Effect of Determining Temperature from the Profiles

The method by which the temperature is determined from the 10.7 cm solar flux was discussed dn Section 2.3.

An alternate method which does not rely on the solar flux



VARIATION IN CALCULATED β_{300} DUE TO DIFFERENT TOP PROFILE ASSUMPTIONS

FIGURE 4

measurements which has been used is to fit a parabola to the under surface of the layer and thus determine the temperature by assuming a relationship between the scale height of atomic oxygen and a parameter of the parabola. One such parameter called SCAT has been introduced by N.B.S. and is published with the electron density profiles.

The thickness parameter SCAT is determined by taking points from the reduced true-height profile and fitting a curve to them. SCAT is given by the relation:

$$SCAT = \sqrt{\frac{n_{max}}{-4\,\frac{(n_3 - n_o)}{(h_o - h_3)^2}}} \qquad (4.2.1)$$

where h_m is the height of point n

n_m is the electron density of point n

and m increases with decreasing altitude.

For a parabolic height distribution:

$$n = n_m \left(1 - \frac{Z^2}{4\,A^2} \right) \qquad (4.2.2)$$

where Z is measured from the layer peak.

$$\frac{\partial n}{\partial Z} = -\frac{Z}{2A^2}\, n_m \,, \qquad (4.2.3)$$

$$\frac{\partial^2 n}{\partial Z^2} = -\frac{1}{2A^2} n_m, \qquad (4.2.4)$$

and SCAT is given by:

$$SCAT = \sqrt{\frac{-n_m}{2\frac{\partial^2 n}{\partial Z^2}}} \quad A \qquad (4.2.5)$$

Nisbet and Quinn (1963) used a relationship for temperature assuming the layer was Chapman like.

For this case,

$$n = n_m \exp \tfrac{1}{2} (1 - z - e^{-z}) \qquad (4.2.6)$$

where $$z = \frac{Z - Z_o}{H} \qquad (4.2.7)$$

At the point where $n = .5\, n_m$

$$Z_m - Z = .707H \qquad (4.2.8)$$

If a parabola is fitted to this Chapman layer at the $0.5\, n_{max}$ point, then:

$$A = 1.846\, H \qquad (4.2.9)$$

and

$$SCAT = .92\, H. \qquad (4.2.10)$$

Temperatures deduced from this relationship were found to be systematically lower than temperatures determined from the 10.7 cm solar flux.

As discussed in Section 1.2 Dungey (1956) calculated theoretical profiles for the shape of the nighttime equilibrium F layer. He presented two coefficients $\underline{a}$ and $\underline{a}'$ relating the scale height of atomic oxygen to parabolas fitted to the layer. $\underline{a}$ related the thickness of a parabola fitted to the $0.5\ n_{max}$ point to the scale height of atomic oxygen and $\underline{a}'$ related the scale height of atomic oxygen to the curvature at the peak of the layer. For the case of zero drift velocity $\underline{a} = 1.63$ H and $\underline{a}' = 1.84$ H.

As the N.B.S. parameter, SCAT, is determined about 30 kilometers below the peak of the layer it may be assumed that, at night, the relationship between $\underline{a}$ and H should be used to determine the ionospheric temperature.

Using the N.B.S. notation, this gives

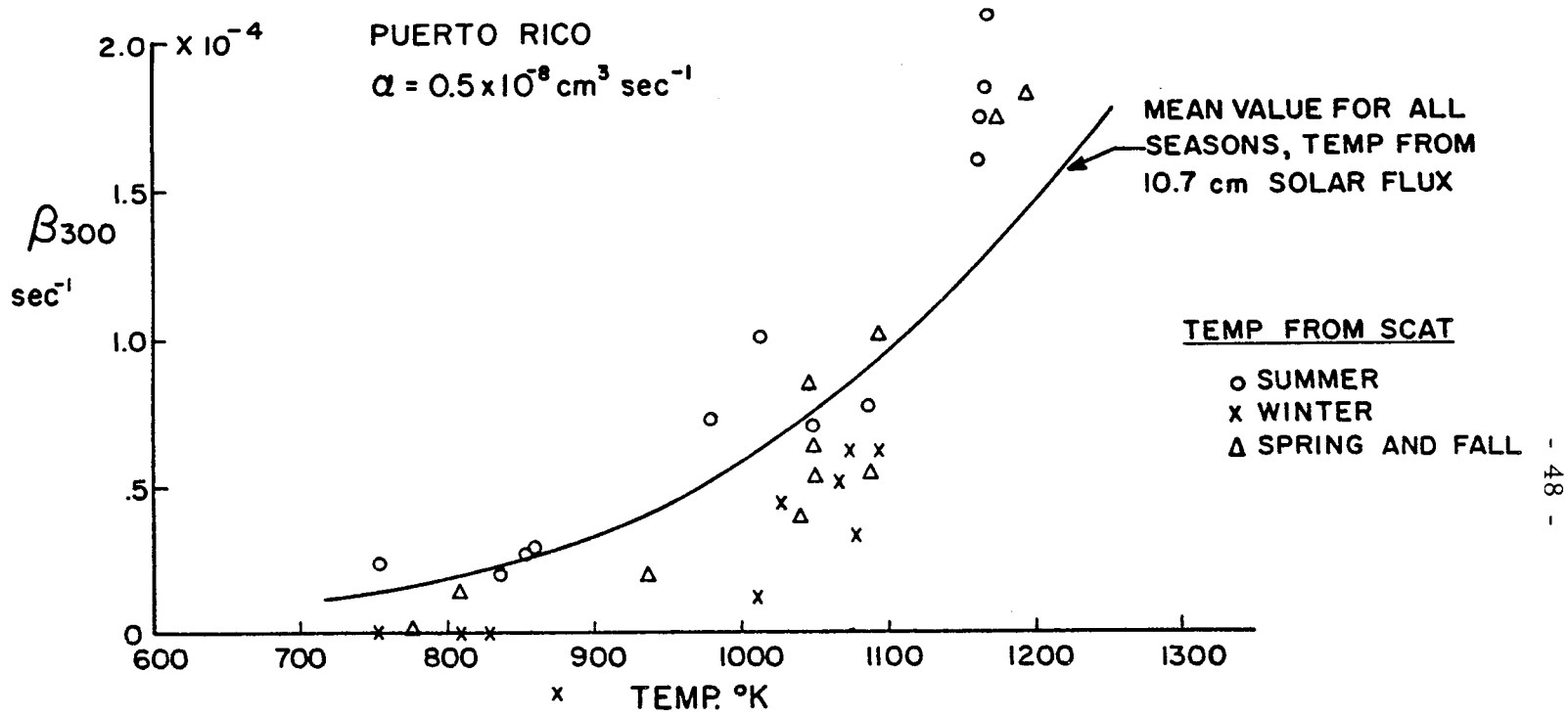
$$SCAT = 0.815H \tag{4.2.11}$$

For example, at 350 kilometers, the temperature determined from SCAT assuming atomic oxygen as the ionizable constituent is given by,

$$T = 21.0\ SCAT \quad {}^{o}K/km \tag{4.2.12}$$

This relationship was used to determine the temperature from the electron density profiles which were used to calculate β_{300} for each month for Puerto Rico. The results are shown in Figure 5.

It is apparent that the temperatures derived from the SCAT values are in general agreement with those determined from the mean monthly 10.7 cm solar flux measurements provided



EFFECT OF DETERMINING TEMPERATURE FROM SCAT

FIGURE 5

the correct relationship between the temperature and SCAT is used for the nighttime profiles.

It is also apparent that the systematic difference between the winter and summer values calculated for β_{300} remains. This difference can thus not be ascribed merely to a temperature difference undetected in the satellite retardation studies from which the 10.7 cm flux relationship was derived.

4.3 The Effect of the Quadratic Loss Coefficient

Probably the most uncertain quantity in this analysis is the value of α, the quadratic loss rate coefficient which appears in equation (3.1.9). A wide range of values has been reported for this coefficient both from ionospheric and laboratory measurements.

A method widely used in the past to determine α involves a study of the decay of ionization during an eclipse of the sun. Nicolet and Hunaerti (1955) concluded that a minimum value of 4×10^{-8} cm^3 sec^{-1} can be deduced from these measurements.

Kasner, Rogers and Biondi (1961) reported a value of $(3.8 \pm 1) \times 10^{-7}$ cm^3 sec^{-1} at 300°K determined by laboratory measurement. However, Swider (1963) quotes Biondi as indicating that this value should be reduced by a factor of 2.

The values quoted above are for the assumption that O_2 is the molecular constituent entering into the charge exchange reaction. If N_2 is the molecular constituent, then a different

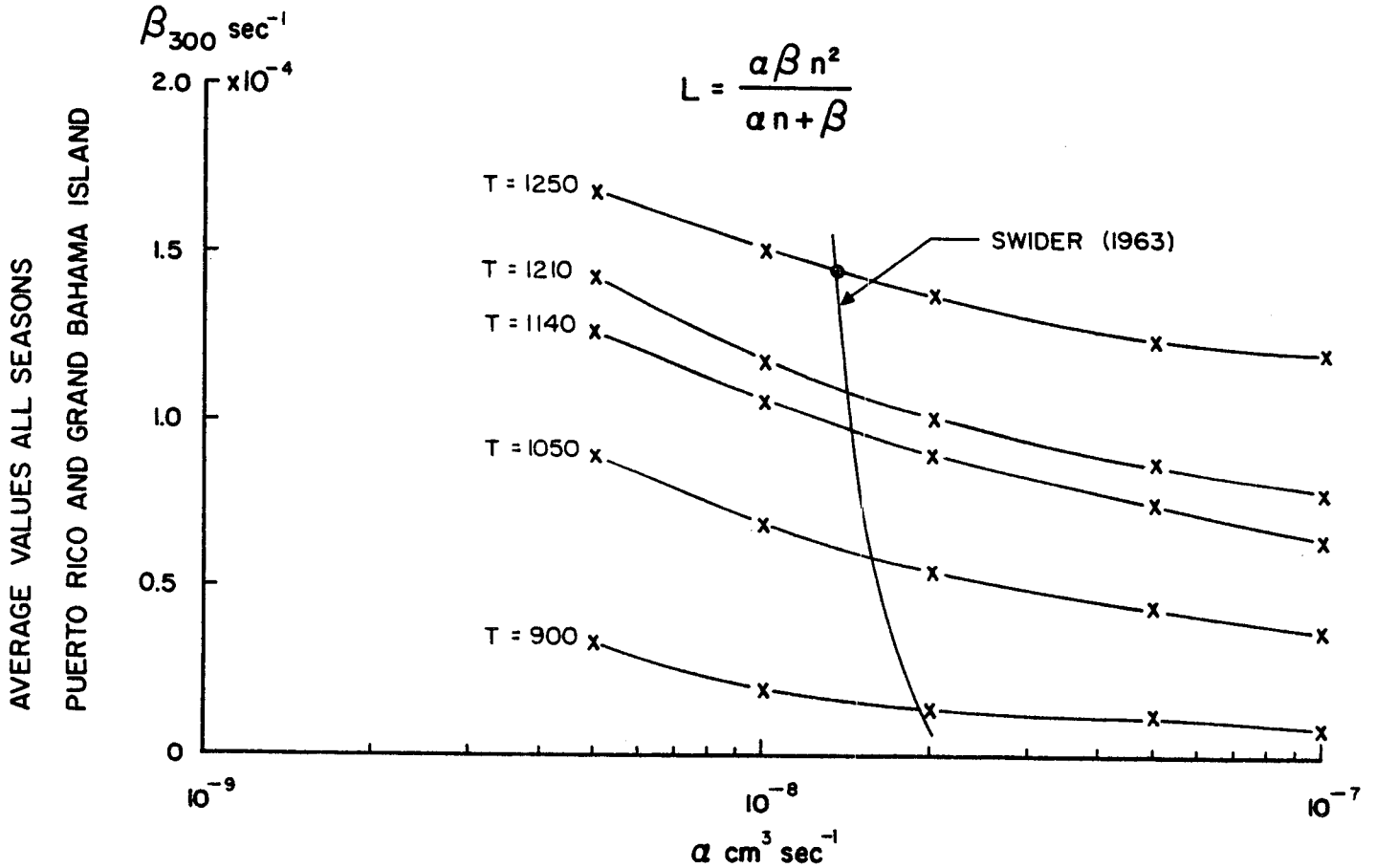
range of values is possible.

In addition to the uncertainty in the value of α at a specified temperature, its temperature dependence is also unknown. Swider (1963) summarized the α variations with temperature which have been used by various ionospheric workers in the past. Expressions involving T^{-1}, $T^{-\frac{1}{2}}$, and $T^{-3/2}$ have been reported. He adopts a value for α of

$$\alpha = 10^{-7} \left(\frac{300}{T}\right)^{3/2} \text{ cm}^3 \text{ sec}^{-1} \qquad (4.3.1)$$

In view of the uncertainty of the quadratic loss rate, β_{300} was calculated for values of α ranging from 0.2×10^{-8} to 4×10^{-7} cm^3 sec^{-1}.

The results of these calculations for five values of temperature are shown in Figure 6. The value given by Swider is plotted on the same graph. The results of this present study indicate that his value is probably an upper bound on α. A value between 0.2×10^{-8} and 1.0×10^{-8} cm^3 sec^{-1} for α has been found to give the most consistent results for the diffusion velocity analysis in this study and because of this, a value of α of 0.5×10^{-8} has been adopted as a best estimate. It should be pointed out, however, that the assumptions about α and about the exact shape of the low lying ionization are extremely interdependent.



VARIATION OF β_{300} WITH THE QUADRATIC LOSS TERM α

FIGURE 6

4.4 Comparison of β_{300} for N_2 and O_2 Molecular Constituents

It was assumed in Section 3.1 that β was proportional to the molecular oxygen density and β_{300} was calculated on the basis of a diffusive equilibrium model for O_2.

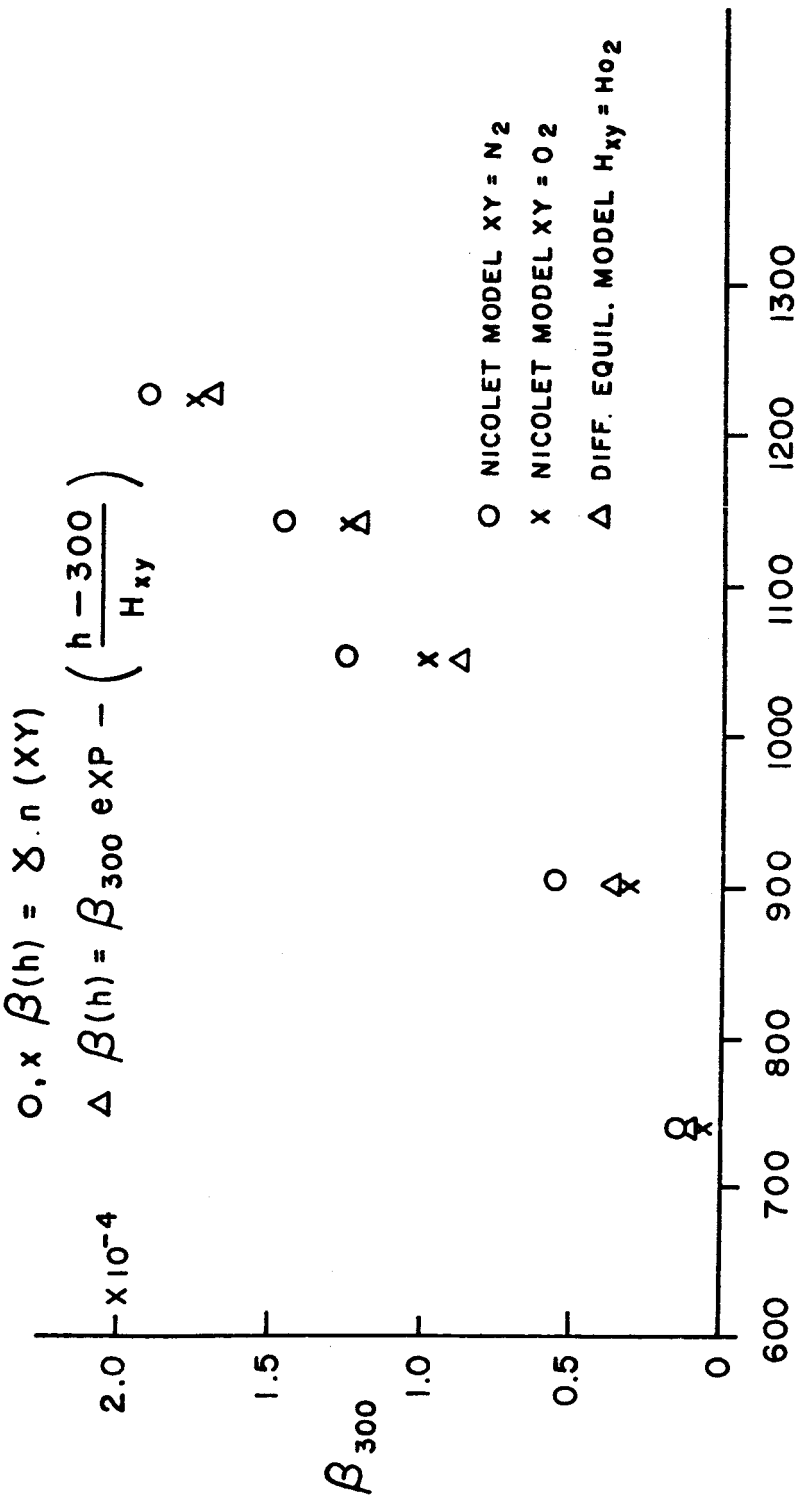
Since it is possible that β is proportional to N_2, an atmospheric model, Nicolet (1961), was used to determine the height distribution for both O_2 and N_2 and β_{300} calculated for each of these distributions. An equation of the form

$$N = N_{01} \exp^{-a(z-300)} + N_{02} \exp^{-b(z-300)} \qquad (4.4.1)$$

was fitted to Nicolet's density data and then used in the computer to determine the height dependence of β for the particular constituent being considered. The height dependence of β under these conditions is given by

$$\frac{\beta}{\beta_{300}} = \frac{1}{N_{01} + N_{02}} \left[N_{01} \exp^{-a(z-300)} + N_{02} \exp^{-b(z-300)} \right] \qquad (4.4.2)$$

The results of these calculations are shown in Figure 7. It is apparent that the model results are in good agreement with the assumption of a diffusive O_2 distribution and it makes little difference whether N_2 or O_2 is assumed to be the molecular



EFFECT OF ASSUMPTIONS ABOUT HEIGHT DEPENDENCY OF β

FIGURE 7

constituent. This of course means that the present analysis cannot determine which molecular constituent enters into the charge exchange reaction with the atomic oxygen ions.

4.5 Low-Lying Ionization

It was noted in section 2.4 that the raw profile data must be corrected for the effects of low-lying ionization, and the method used to implement the correction was described.

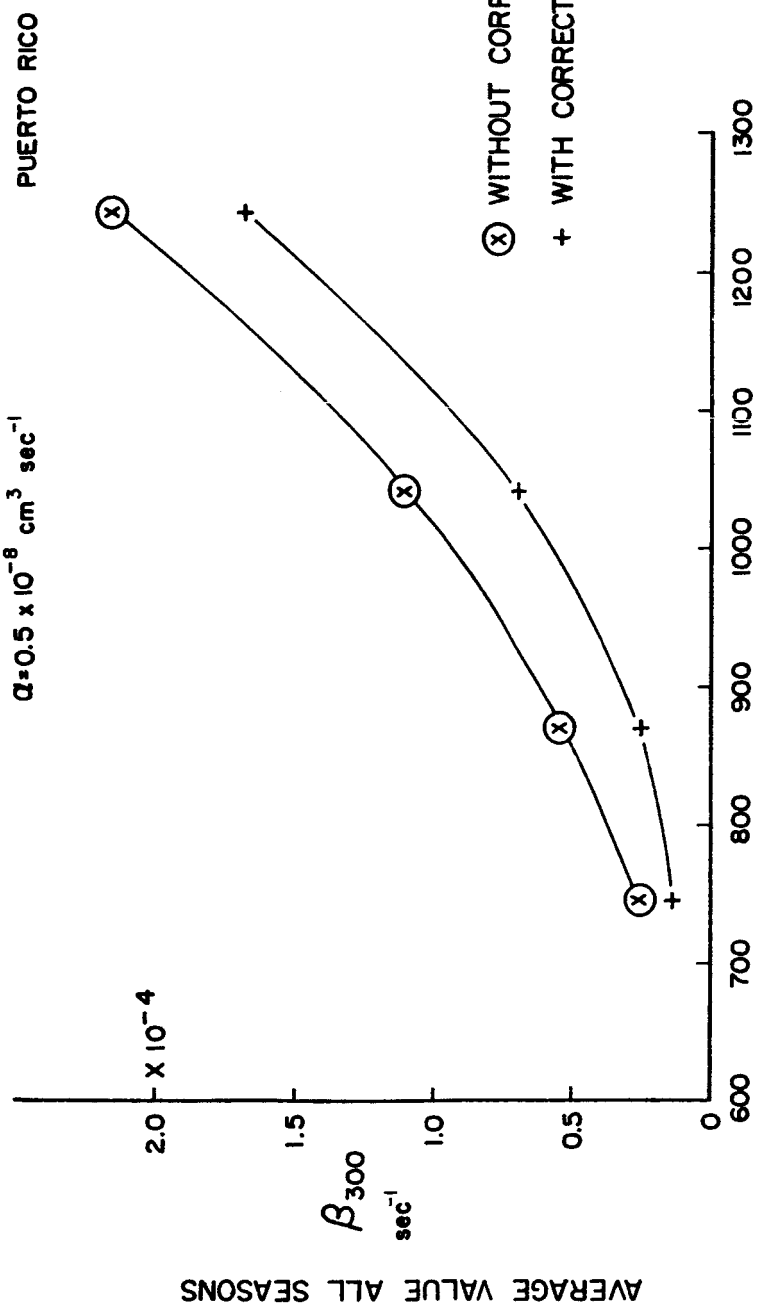
To investigate the effect of neglecting this low-lying ionization, β_{300} was calculated for the two cases with and without the correction. The results are shown in Figure 8.

Clearly the neglect of low-lying ionization can lead to significant errors in the value for the recombination coefficient. This further emphasizes the need for accurate measurement and reduction techniques to determine nighttime electron density profiles requiring a minimum of correction.

4.6 Production Due to Protonosphere Coupling

Iwai and Outsu (1958), Rivault and Corcuff (1960) and Allcock (1960) have reported diurnal changes of the order of 30% in the total content of the protonosphere as observed from Whistler measurements. Carpenter (1962) on the other hand reports a significant number of cases where no variation occurred during the night. He does find a 20 to 25 percent reduction in density from early 1958 to early 1961.

The diurnal change in protonosphere content represents a source of production for atomic oxygen ions at night if it is



EFFECT OF NEGLECTING LOW-LYING IONIZATION

FIGURE 8

assumed that the protons move down the field lines and produce atomic oxygen ions by means of charge exchange. The effect of such production on the magnitude of the recombination coefficient was examined as follows.

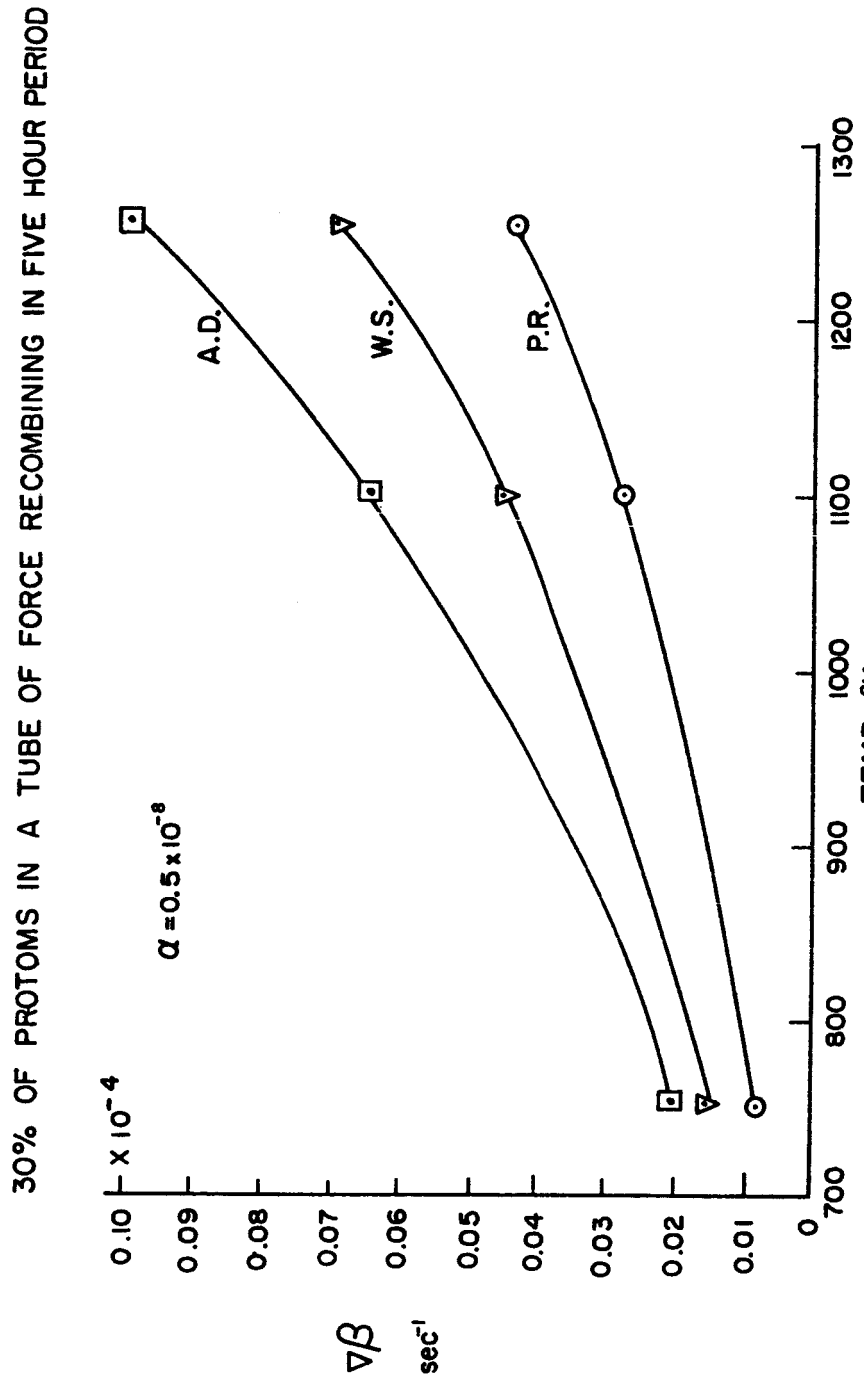
An upper limit on the decrease of protonosphere density was taken as 30 percent in a 12 hour period. To utilize this decay as a production mechanism for atomic oxygen ions, the total proton content in a magnetic tube between 1000 km and the geomagnetic equator was calculated on the basis of the protonosphere densities presented by Smith (1961). It was assumed that the decay for the period 2300 hours to 0300 hours resulted in an equal number of atomic oxygen ions being produced.

The results of this calculation for three stations are shown in Figure 9.

It is apparent that if protonosphere coupling does act as a source of production at night its effect is quite small. This could however be important in cases at low temperatures when the recombination coefficient becomes very small. Its effect in such cases is further discussed in Section 5.3.

4.7 The Effect of a Change in Te/Ti

A diurnal change in the electron-ion temperature ratio is a factor which could effect β_{300} in the same way as nighttime production because of the assumptions about the top of the layer made in this analysis.



CHANGE IN β_{300} DUE TO PRODUCTION AT NIGHT BY PROTON CHARGE EXCHANGE WITH ATOMIC OXYGEN

FIGURE 9

If it is assumed that this ratio has not yet decreased to unity at the time when the analysis begins, 2200 hours and that it does decrease to unity during the five hour period, 2200 hours to 0300 hours, the top content change due to this change in the profile may be estimated and its effect on β_{300} calculated. A decrease in Te/Ti from 1.5 to 1 would produce about the equivalent of 20 percent of the top content at 2200 hours added to the normal change in top content.

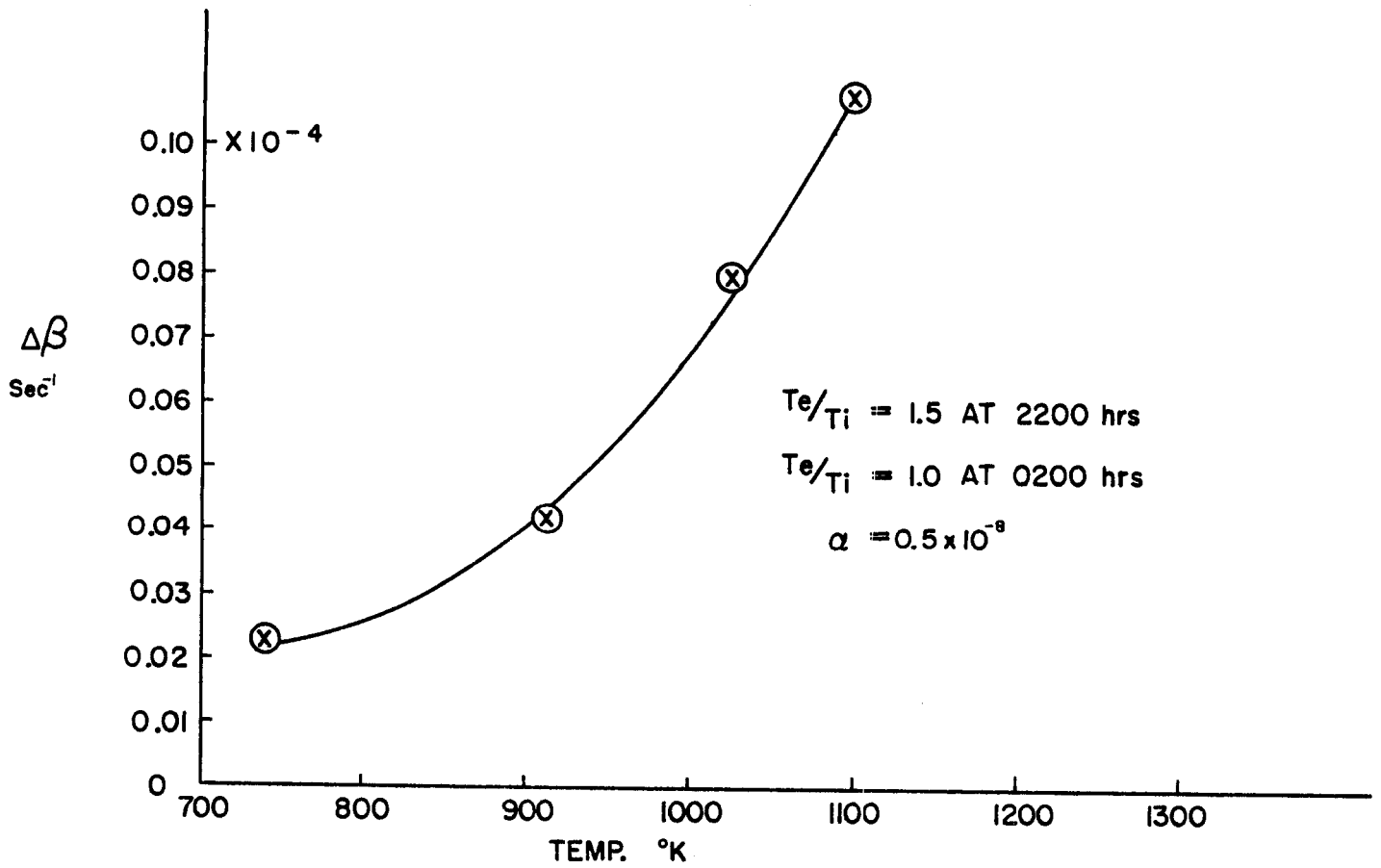
If this effect is present it would result in a change in β_{300} as illustrated in Figure 10.

As is the case for proton production, a change in the Te/Ti ratio during the night could be important when the magnitude of β_{300} is very small and it is also further considered in Section 5.3.

4.8 β_{300} For Individual Days

The recombination coefficient calculations discussed in previous sections have all been based on monthly mean electron density profiles derived by means of averaging the daily profiles for days when $K_p < 4.5$.

In order to compare these results with the recombination coefficient for each day, β_{300} was calculated using daily profiles for Puerto Rico published by C.R.P.L.. Two winter and two summer months were selected and β_{300} calculated for each day when $K_p < 4.5$.



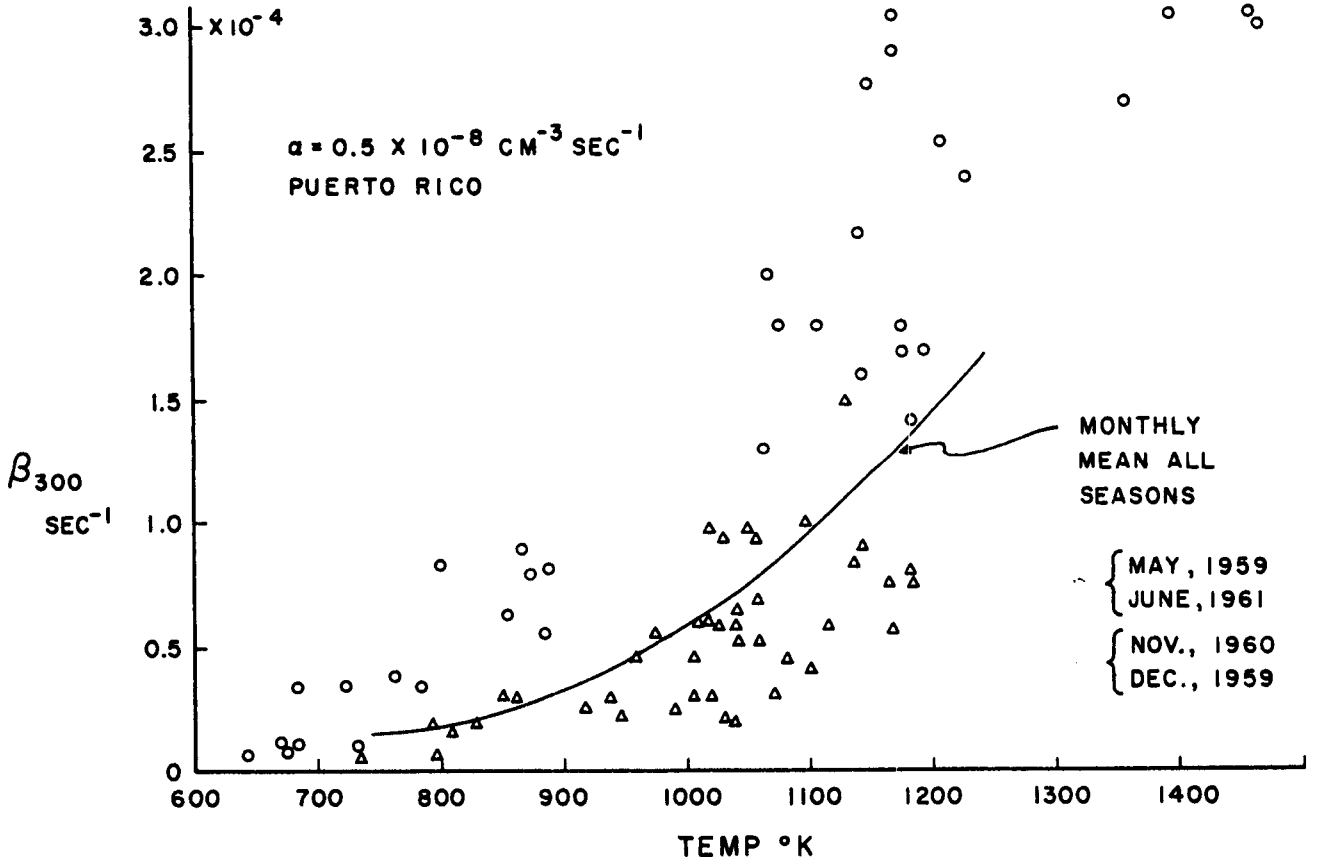
EFFECT OF Te/Ti RATIO CHANGING ON CALCULATED β_{300}

FIGURE 10

Because the Harris and Priester temperature formula does not take into account the "27 day" solar oscillations, the calculations were done using both the Harris and Priester formula and Jacchia's temperature relationship, equation (2.3.4).

The use of these different temperature relationships did not result in any significant difference in the calculated value of β_{300}. The results are shown in Figure 11. The solid line on the graph indicates the mean of the calculated values of β_{300} for the monthly mean profiles.

It is apparent that the daily values are in general agreement with the value of β_{300} based on monthly mean profiles. The seasonal effect is again clearly demonstrated in that the summer values are high and the winter values low.



β_{300} FOR INDIVIDUAL NIGHTS FOR 2 SUMMER AND 2 WINTER MONTHS

FIGURE II

5. RESULTS OF THE ANALYSIS

5.1 Best Estimate for β_{300}

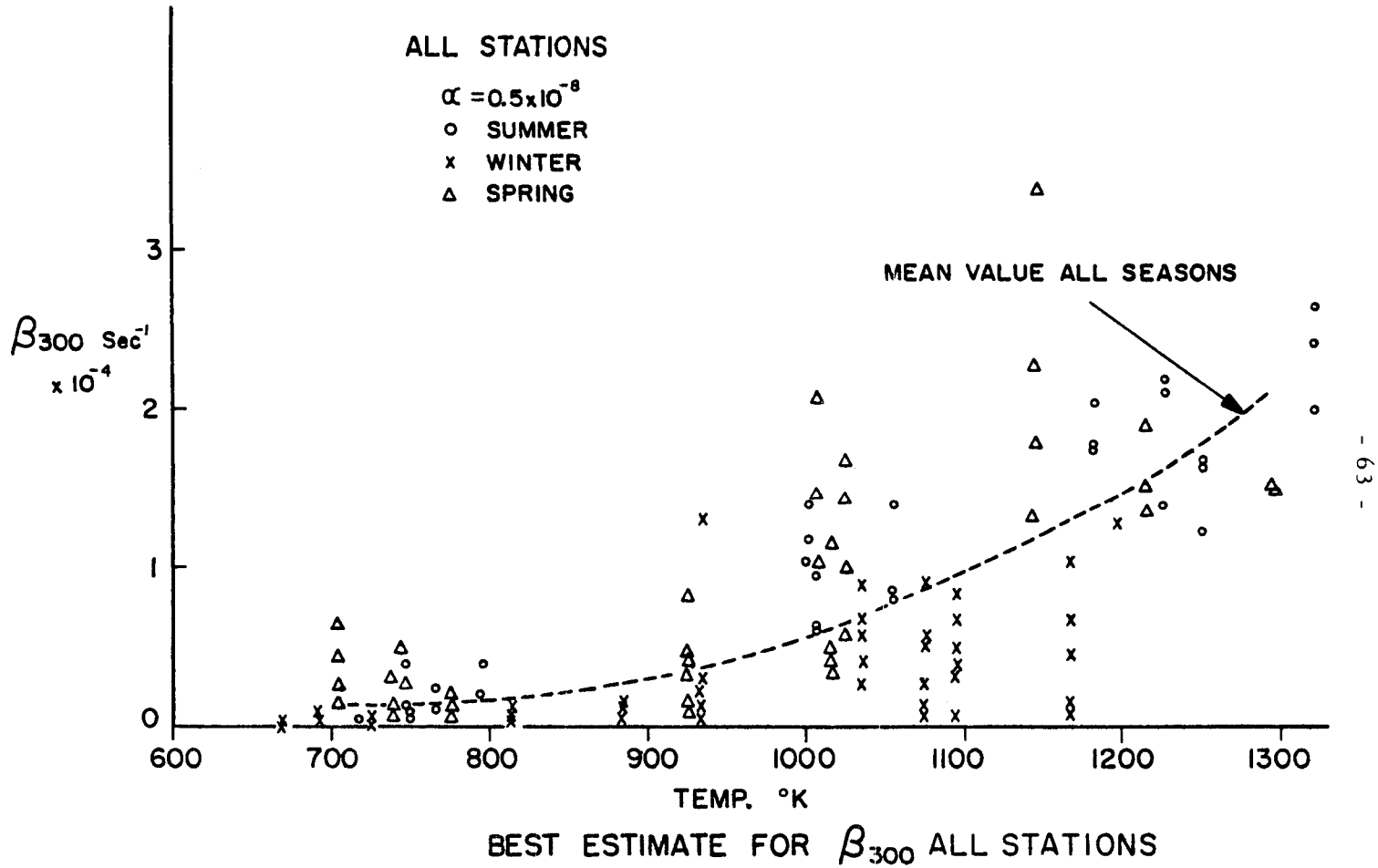
Taking into account the various effects described in Section 4, it is suggested that the best estimate for β_{300} is as shown in Figure 12.

This "best estimate" for β_{300} was determined under the following conditions: a value of 0.5×10^{-8} cm^3 sec^{-1} for the quadratic loss coefficient, α, was used; the profiles were corrected for low-lying ionization; the top profiles described in Section 2.2 were used; and no effective production, due to proton charge exchange or a change in Te/Ti was included.

Figures 13 through 18 show β_{300} for each of the stations used in the analysis. Fewer months were analyzed for the higher latitude stations because for these stations, in summer, the upper layer is sunlit throughout the night. The general agreement for all stations, presence of the seasonal effect and lack of any latitude dependence is very striking.

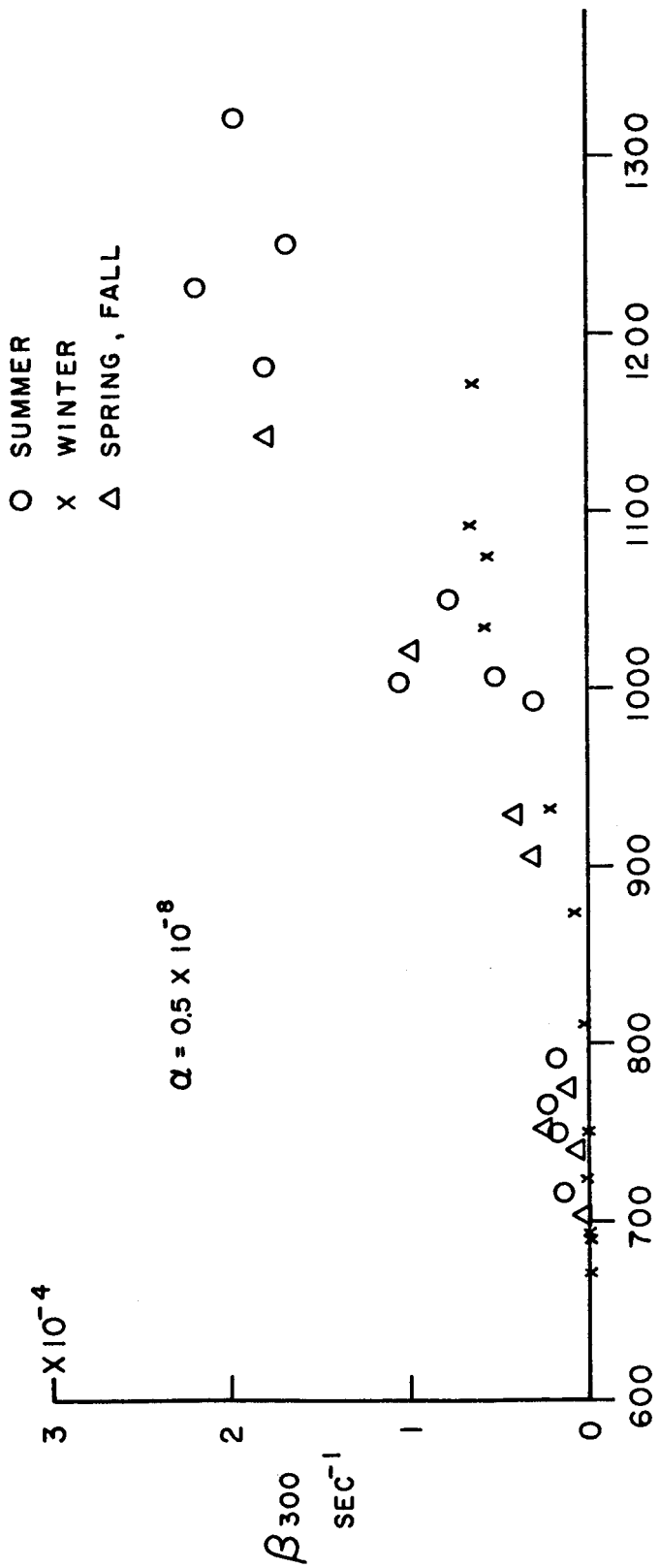
5.2 Best Estimate for the Diffusion Coefficient

A method by which the diffusion coefficient for atomic oxygen ions in the upper F region could be determined was described in Section 3.2. This method was applied to the available electron density profile data and an estimate of the diffusion coefficient determined. The results of this analysis are shown in Figure 19 where the plotted points represent smoothed data



BEST ESTIMATE FOR β_{300} ALL STATIONS

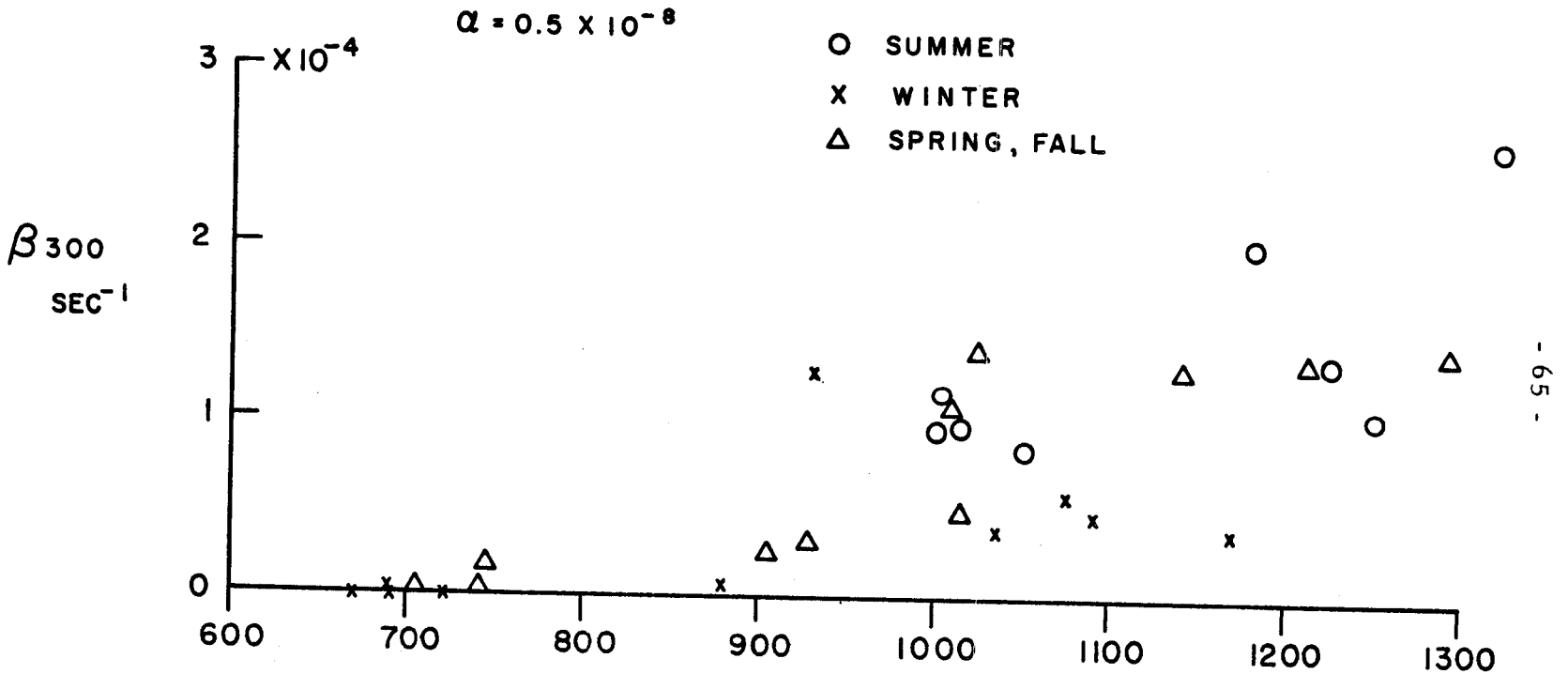
FIGURE 12



BEST ESTIMATE FOR β_{300}

PUERTO RICO

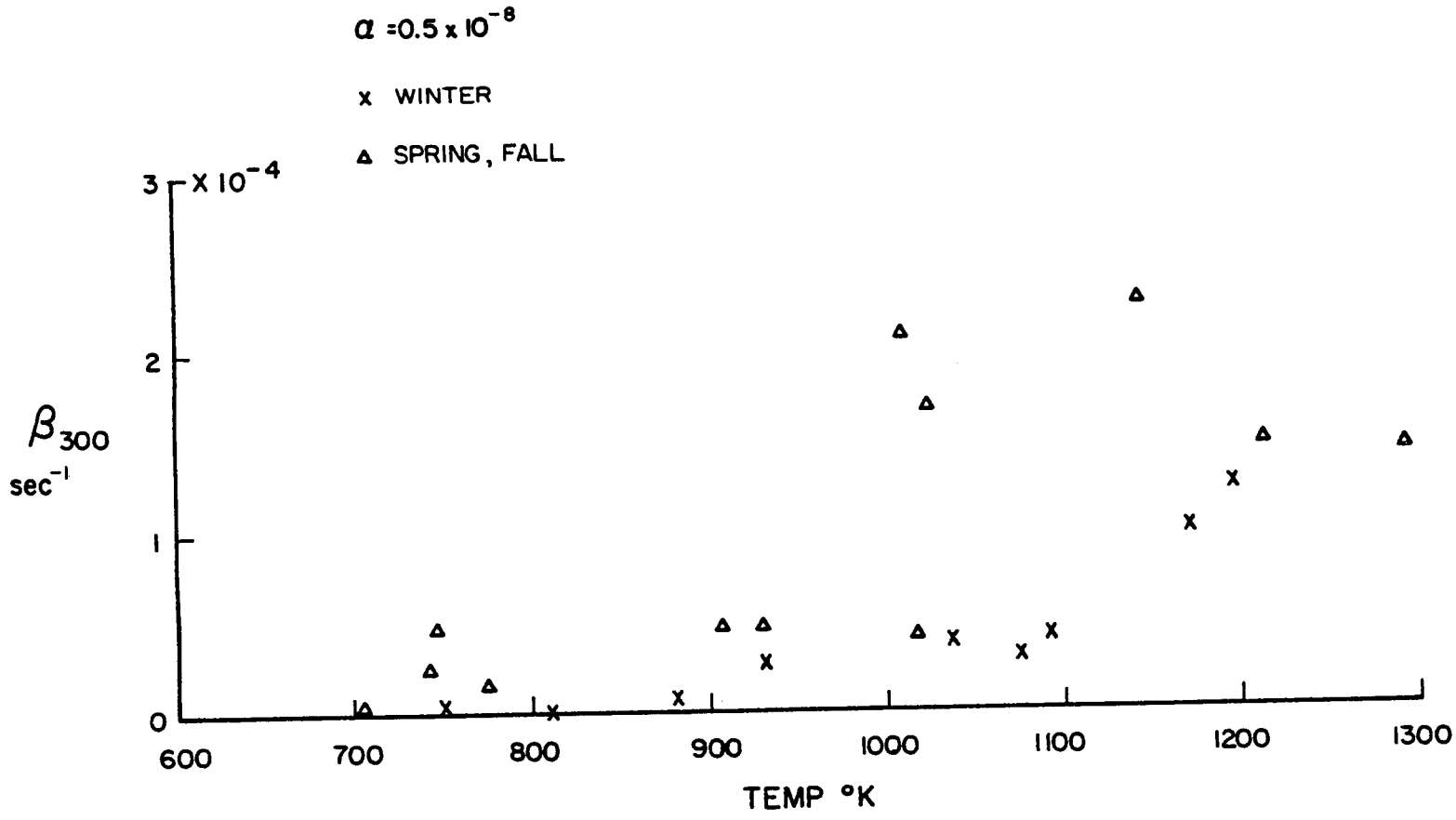
FIGURE 13



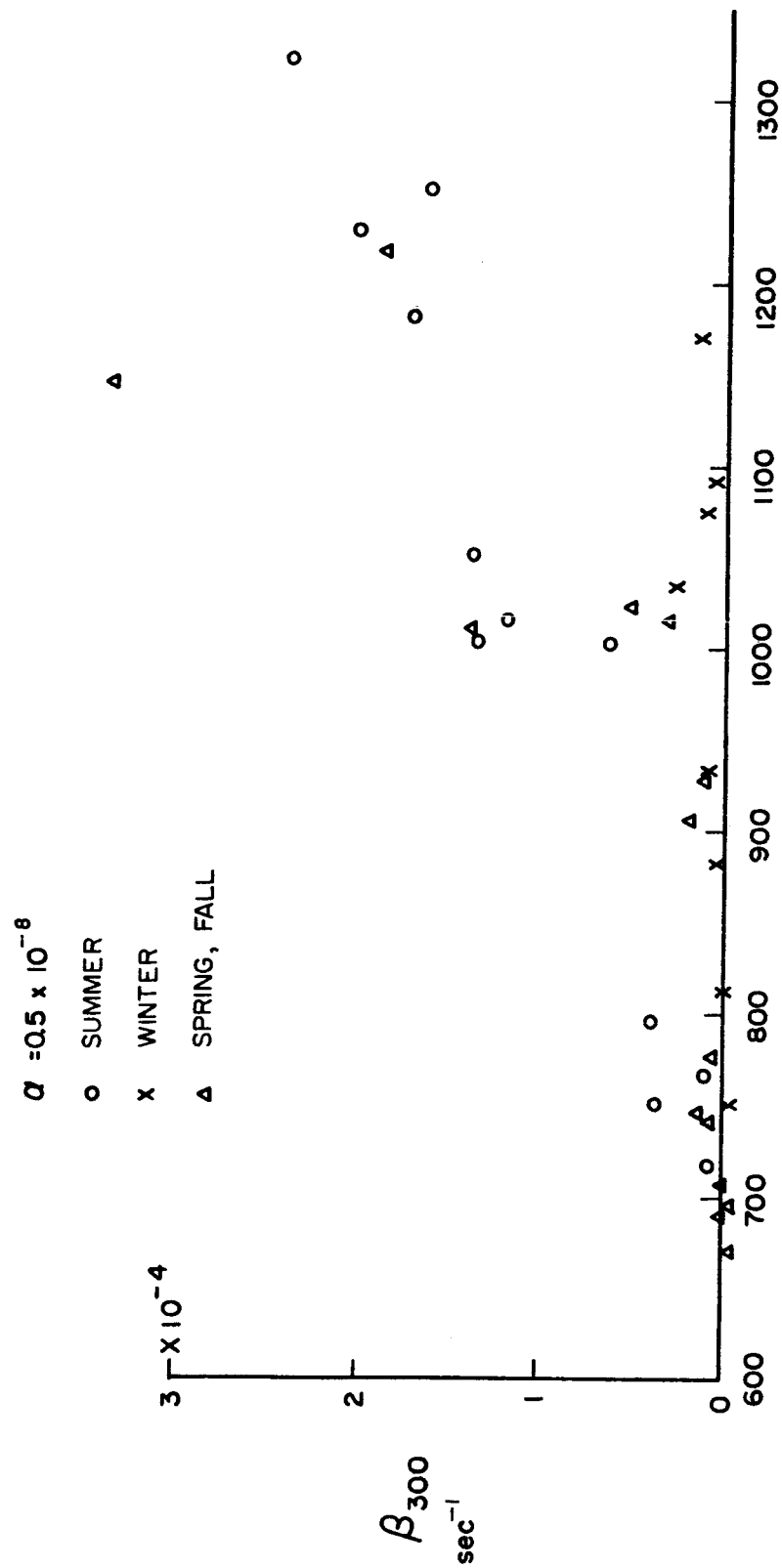
BEST ESTIMATE FOR β_{300}

GRAND BAHAMA ISLAND

FIGURE 14

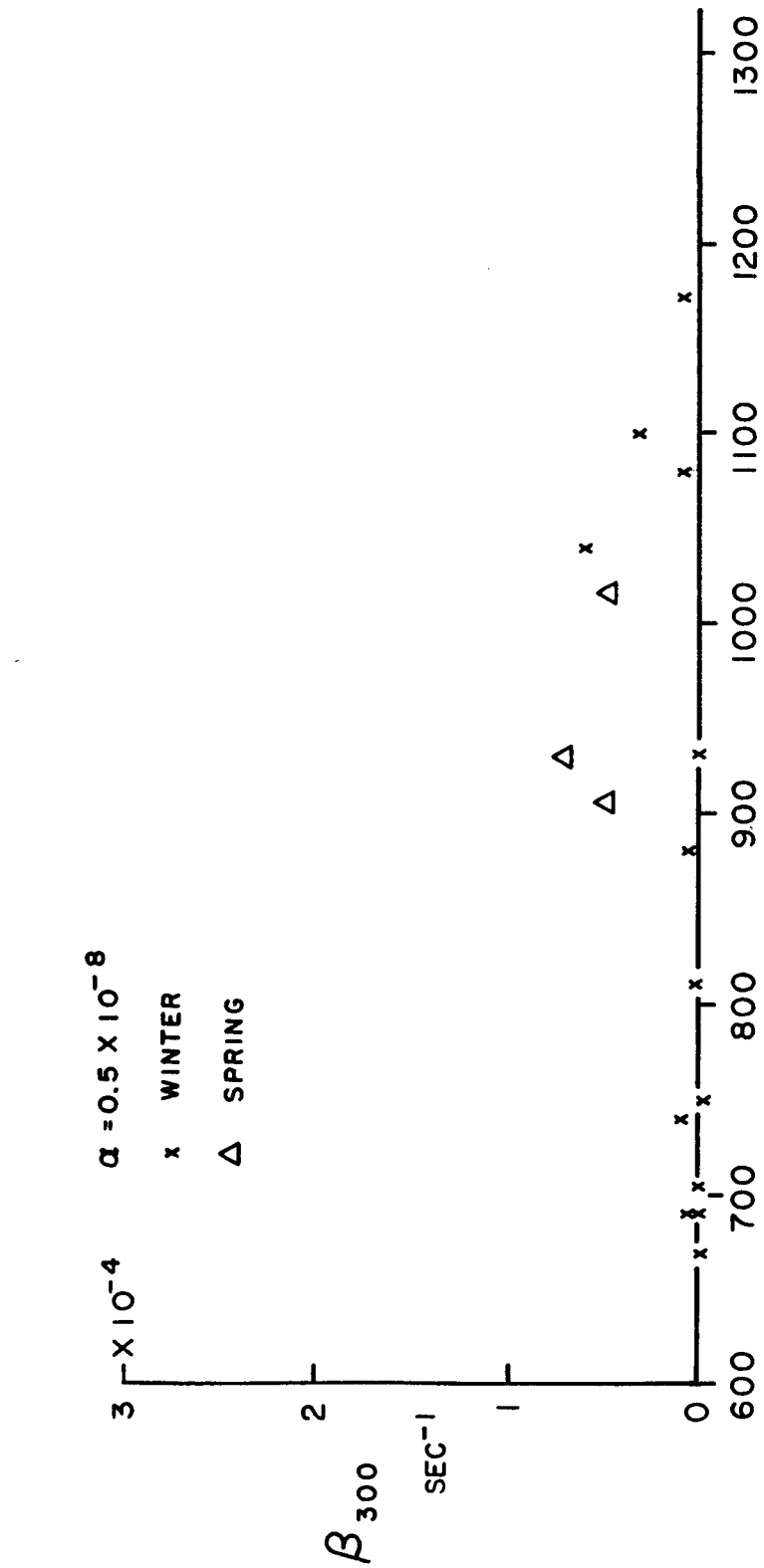


BEST ESTIMATE β_{300}
FORT MONMOUTH, N.J.
FIGURE 15



BEST ESTIMATE β_{300}
WHITE SANDS, N.M.

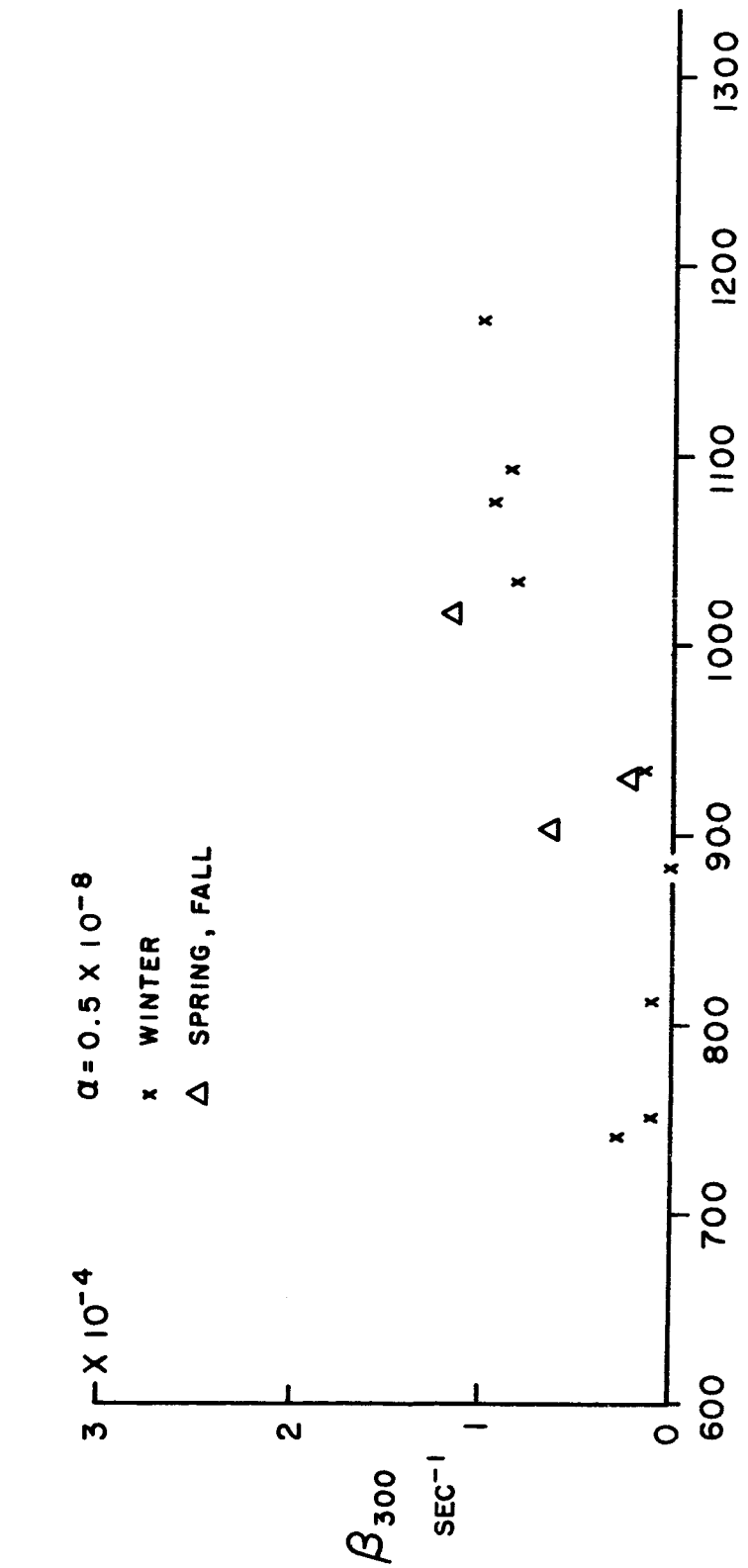
FIGURE 16



BEST ESTIMATE FOR β_{300}

ADAK, ALASKA

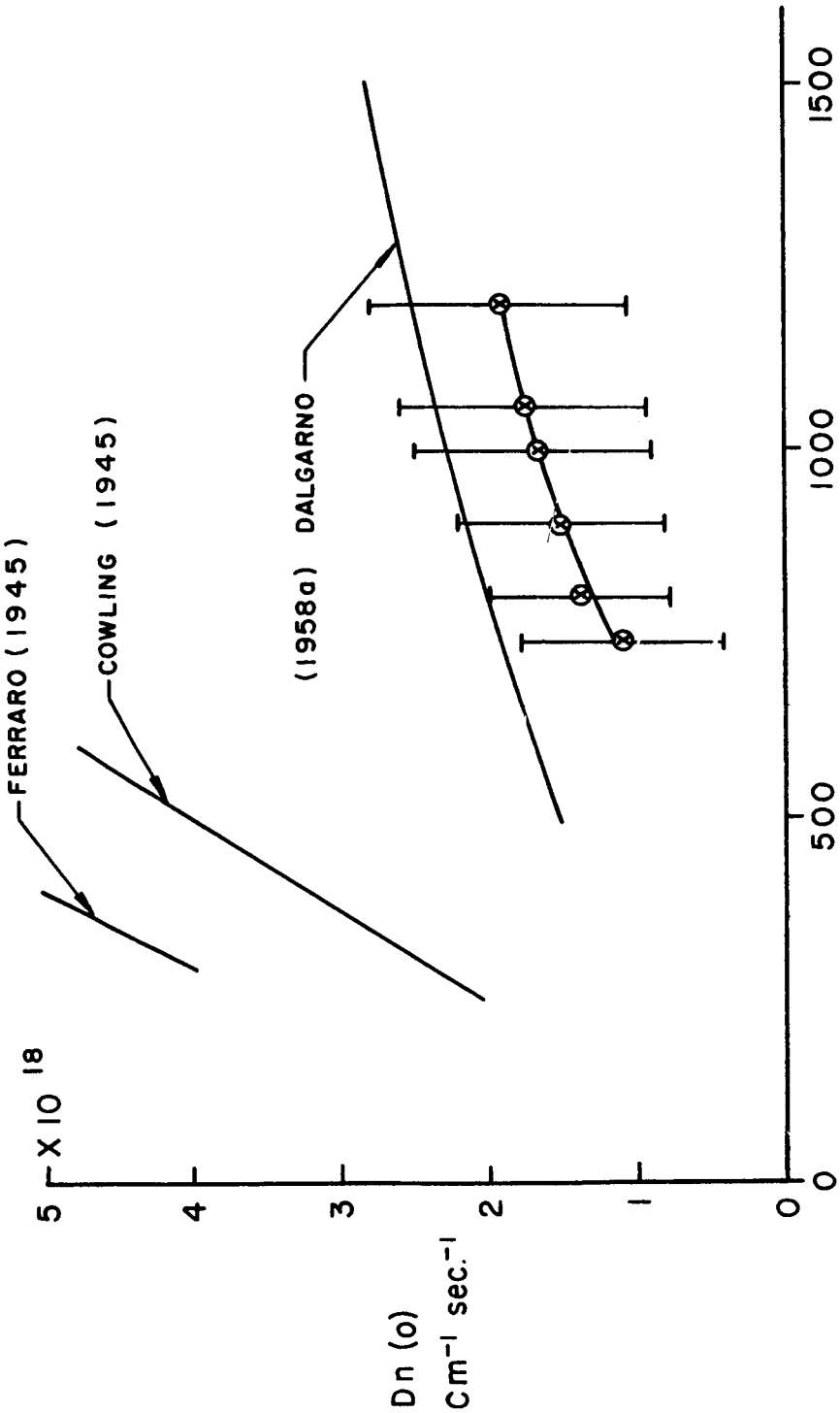
FIGURE 17



BEST ESTIMATE FOR β_{300}

NEW FOUNDLAND

FIGURE 18



NORMALIZED DIFFUSION COEFFICIENT

FIGURE 19

and the error bars show the standard deviation. Values suggested by previous workers are also shown on the graph, and it is clear that the best agreement is with Dalgarno's (1958) values although the values from this study are somewhat lower. The values of D n(M) determined from the smoothed data show roughly a $T^{0.8}$ temperature dependence, with the expression for the diffusion coefficient given by:

$$D = \frac{(6.7 \pm 3) \times 10^{-5} \, T^{0.8}}{n(M)} \sin^2 I \qquad (5.2.1)$$

However, differences between the measurements at different temperatures were not sufficient to make the difference between 0.8 and 0.5 statistically significant and the theoretically probable value of 0.5 was used to repeat the analysis for the Puerto Rico station. From this latter analysis and using the expression for the diffusion coefficient suggested by Shimazaki (1957), D is given by:

$$D = \frac{b \sqrt{T}}{n(M)} \sin^2 I \qquad (5.2.2)$$

Assuming the neutral constituent n(M) to be atomic oxygen and using Nicolet's model, b may be determined. The value for b determined from the present study is presented in Table 2 along with values due to previous workers.

Table 2

Comparison of Diffusion Constants

$\frac{D\, n(0)}{\sqrt{T}}$	Author
4.5×10^{17}	Shimazaki (1957)
3.02×10^{17}	Ferraro (1945)
2.07×10^{17}	Cowling (1945)
0.73×10^{17}	Dalgarno (1958)
$(0.50 \pm 0.3) \times 10^{17}$	Present Study

It is again apparent that the best agreement is with the value presented by Dalgarno.

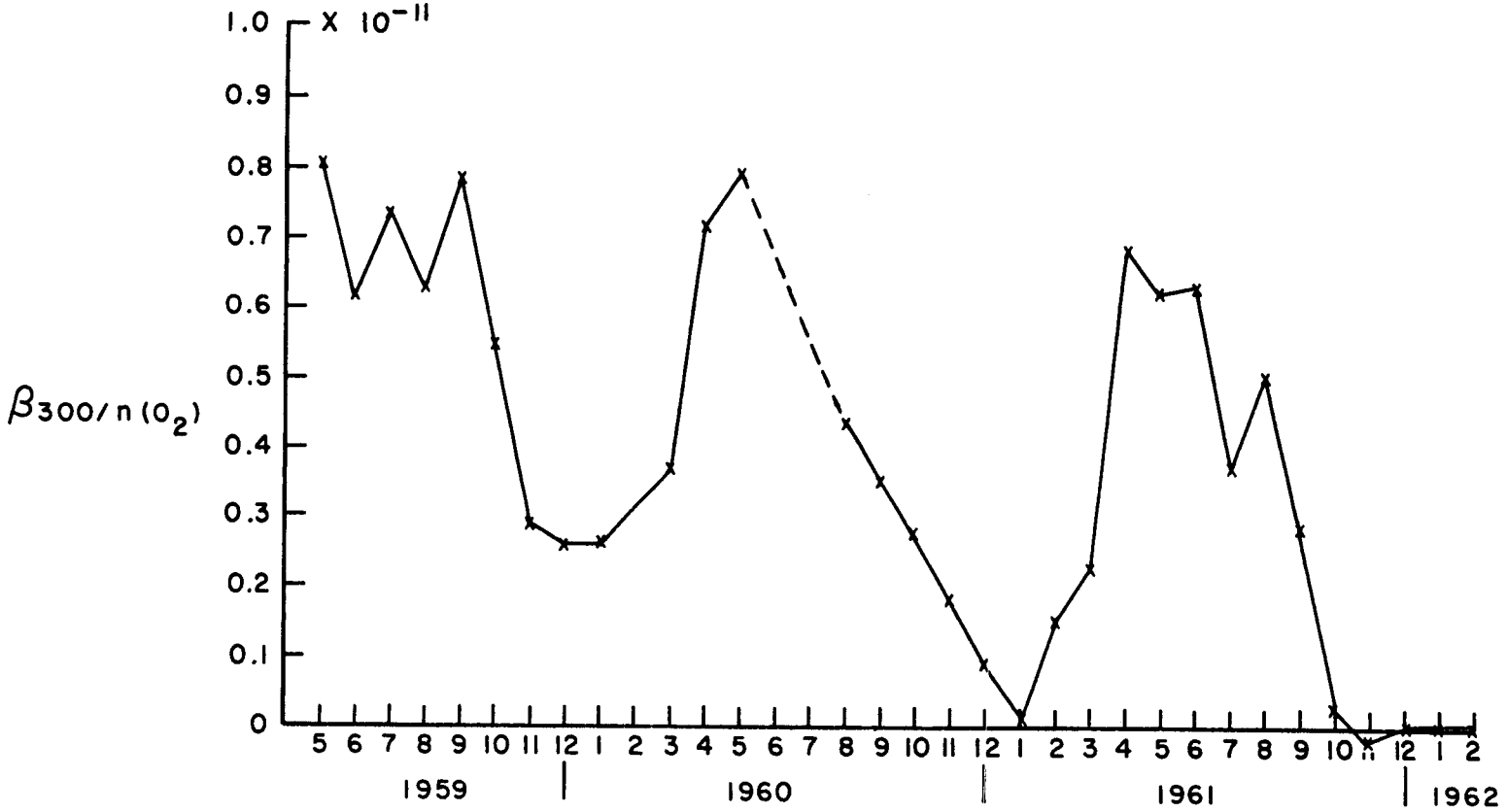
From the results of this study the following value for the diffusion coefficient is suggested:

$$D = \frac{(0.5 \pm 0.3) \times 10^{17}}{n(M)} \quad \sqrt{T} \quad \sin^2 I$$

5.3 Seasonal Effects

It has been pointed out that a strong seasonal dependence for β_{300} has been noted. If the ratio $\beta_{300}/n(O_2)$ is calculated for each month, using Nicolet's model, and the values plotted on a yearly basis this seasonal effect becomes quite clear. A plot of this ratio is shown in Figure 20.

The seasonal effect is quite apparent, but the negative part of the curve for the winter of 1961 is somewhat disturbing.



RATIO OF CALCULATED β 300 VALUES TO $n(O_2)$ DENSITIES USING MODEL WITH NO SEASONAL VARIATION: $\alpha = 0.5 \times 10^{-8}$ PUERTO RICO

FIGURE 20

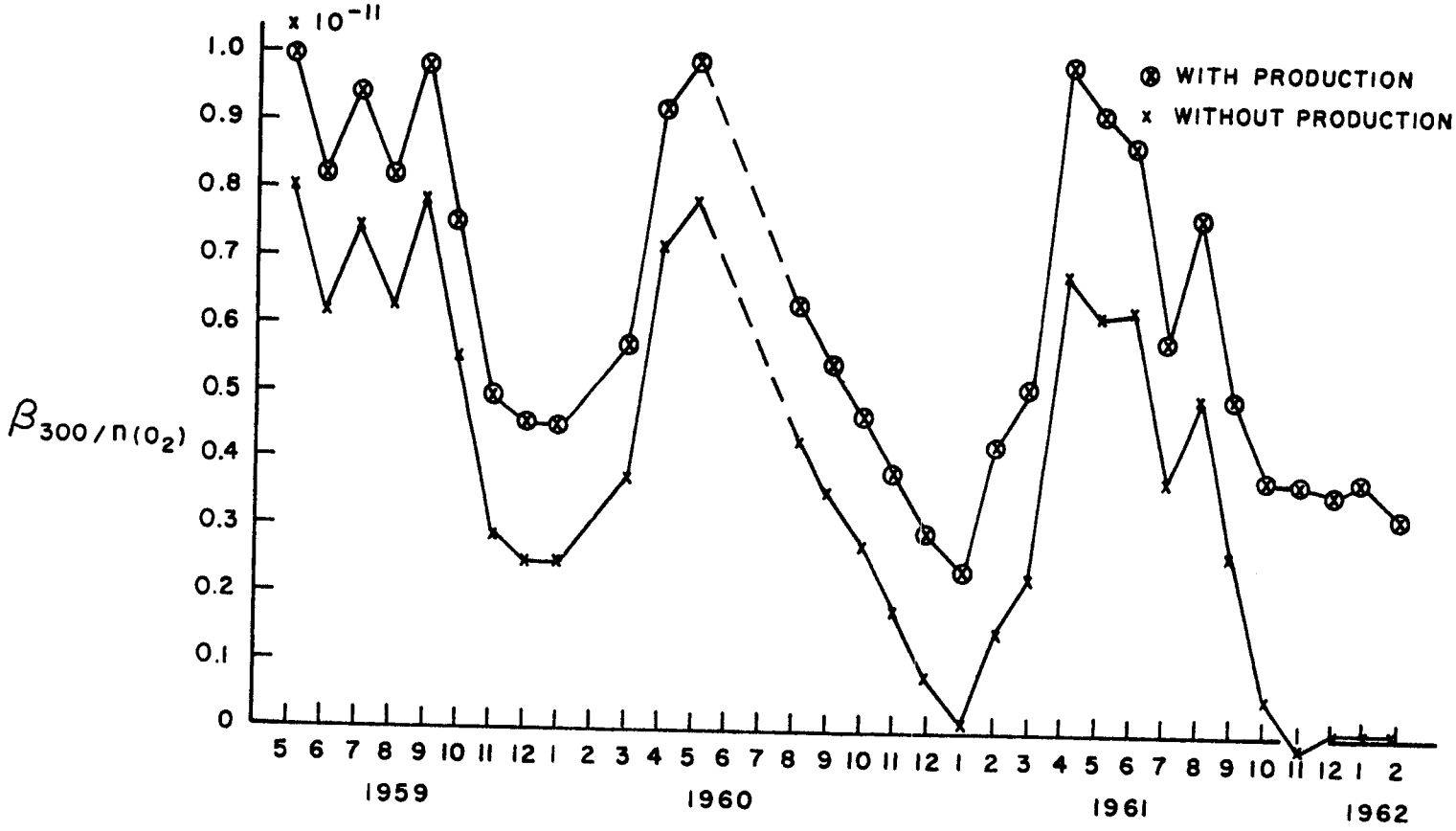
Negative values for the ratio are not acceptable and they suggest that a quasi production mechanism must be operative at night.

Two sources of such production, either protonosphere coupling or a change in Te/Ti, were suggested in Sections 4.6 and 4.7. The effect of such production on the ratio $\beta_{300}/n(0_2)$ has been examined.

An upper bound on the effect of protonosphere coupling can be found by assuming that the proton content undergoes a 100 percent change during the 5 hour period and that the entire change is a source of production for atomic oxygen ions. If these assumptions are made, the seasonal curve will be modified as shown in Figure 21. Even this amount of production is not sufficient to produce effects as large as the seasonal anomaly.

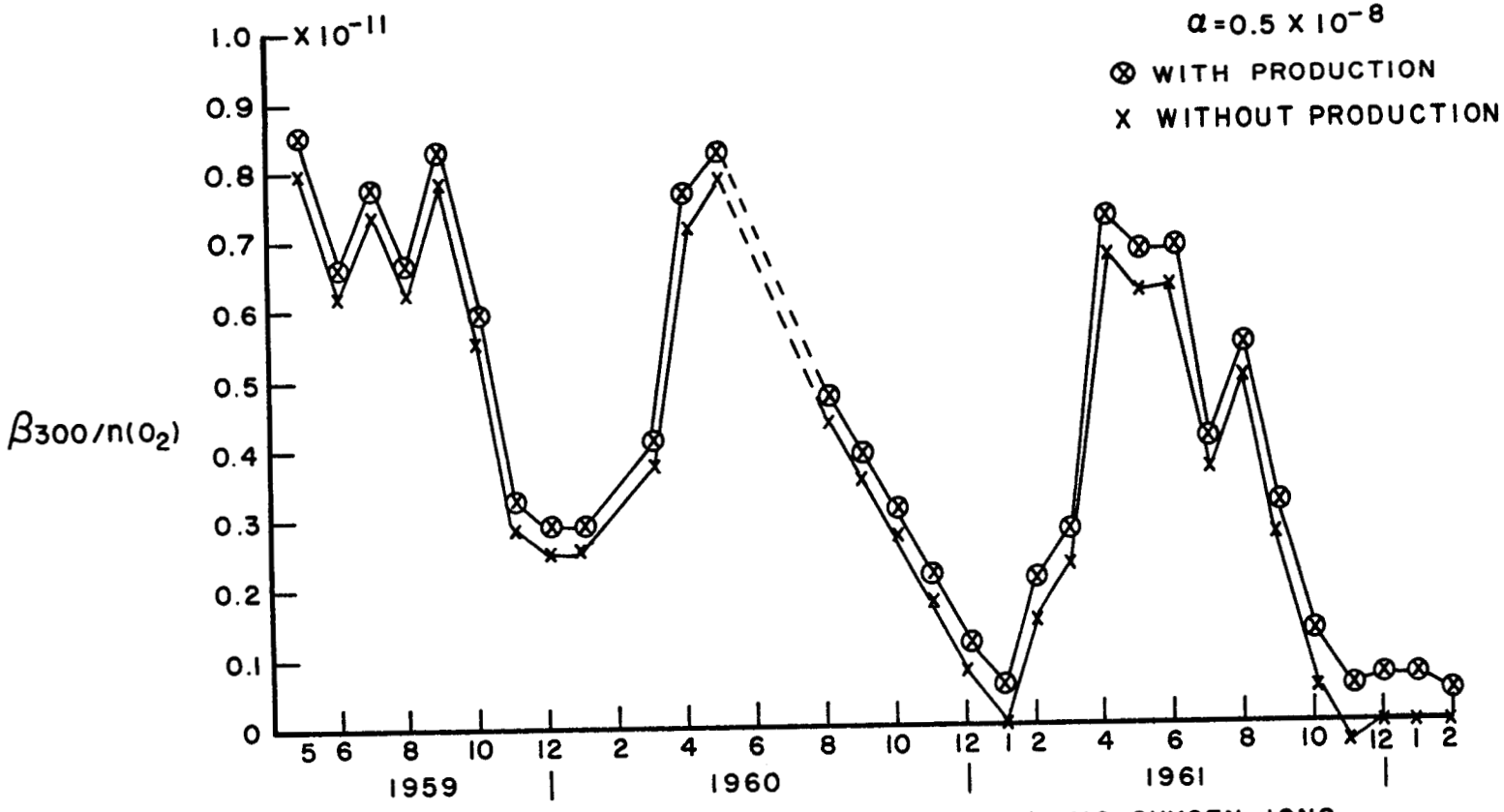
A more reasonable upper estimate of a 30 percent change in the proton content was made and this result is shown in Figure 22. This amount of production would be adequate to ensure that in winter for low sunspot conditions the calculated recombination coefficients are positive for all months, but would not greatly affect values for high sunspot levels.

Values of Te/Ti of about 3 during the day have been suggested and since no good measurements of the relaxation time have been made a change in this ratio during the night has also been considered. A reduction of Te/Ti from 1.5 at 2200



UPPER BOUND STUDY OF PRODUCTION OF ATOMIC OXYGEN IONS BY 100 % PROTONOSPHERE CONTENT FROM 2200 TO 0300 HOURS

FIGURE 21



HIGH ESTIMATE STUDY OF PRODUCTION OF ATOMIC OXYGEN IONS
CORRESPONDING TO 30% OF PROTONOSPHERE CONTENT
FROM 2200 TO 0300 HRS.

PUERTO RICO

FIGURE 22

hours to 1.0 at 0300 hours was used to modify the top profiles used for the $\beta_{300}/n(O_2)$ calculation and the results are shown in Figure 23. It is apparent that an assumed decrease in Te/Ti during the night can remove the very low and negative values calculated for winter low sunspot conditions, while producing a much smaller percentage change for high sunspot conditions.

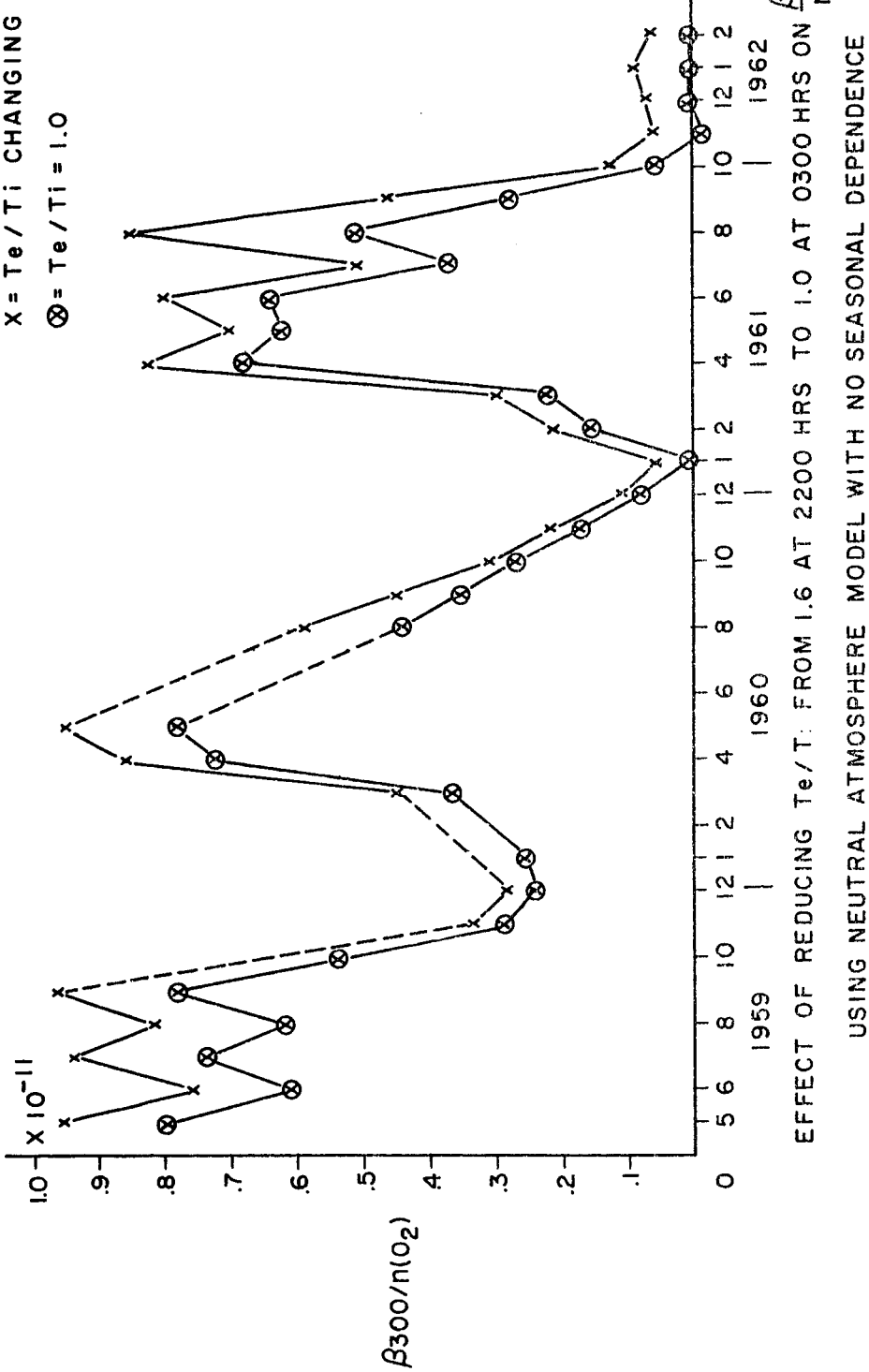
In all cases studied the seasonal effect was very marked and possible seasonal changes in the two production mechanisms examined were insufficient to account for the observations.

The seasonal dependence of β_{300} must reflect a seasonal change in either the O_2 or N_2 density at 300 kilometers. When the values of β_{300} are divided by neutral atmosphere molecular densities for models with no seasonal variation the calculated values show a seasonal variation.

5.4 Drift Velocity

In the process of determining the diffusion coefficient it was first necessary to remove the steady or drift component from the total velocity by the technique described in Section 3.2. Thus a determination of the drift velocity was also made during the diffusion coefficient analysis.

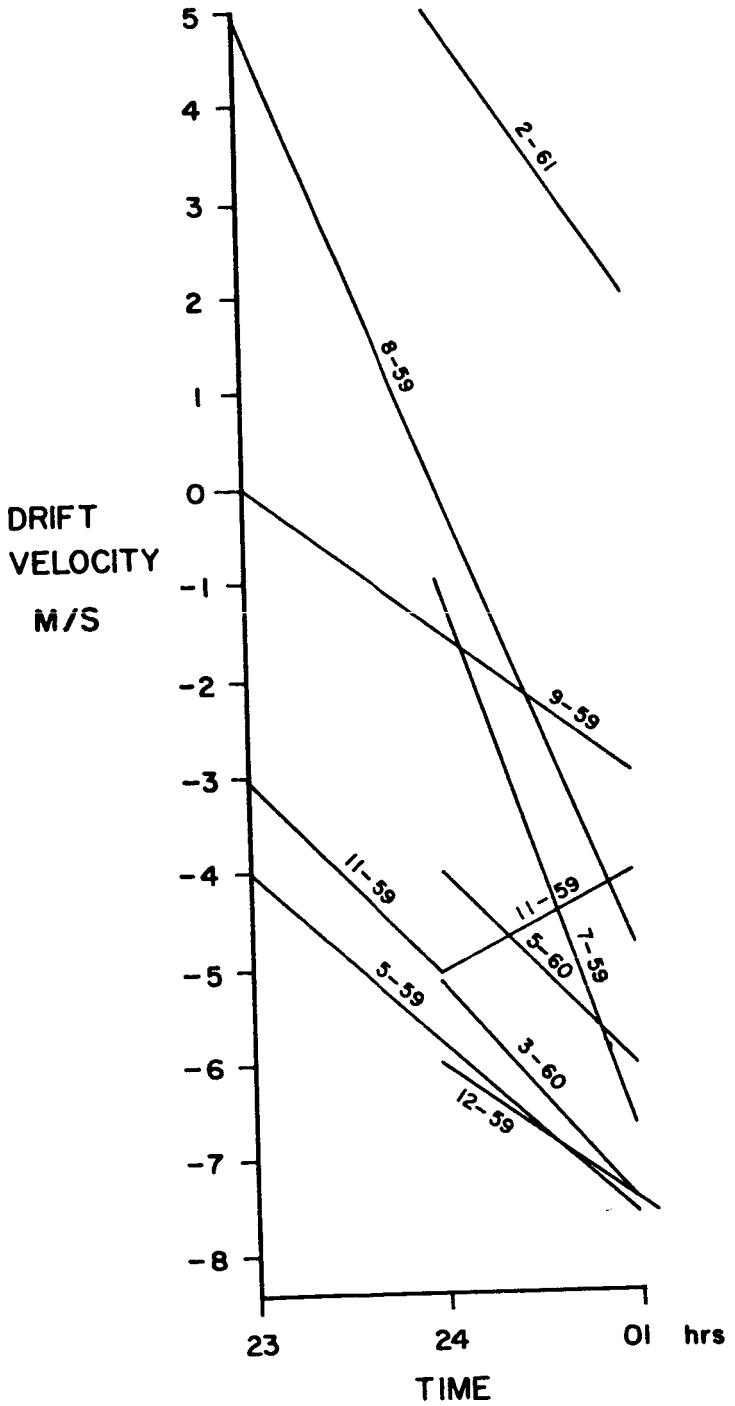
Figure 24 shows a plot of drift velocities for the period 2300 hours to 0200 hours for the Puerto Rico station. These typical results were obtained by integrating equation 3.2.1 in a downward direction and finding the velocity at the lower height



EFFECT OF REDUCING Te/Ti FROM 1.6 AT 2200 HRS TO 1.0 AT 0300 HRS ON $\frac{\beta 300}{n(O_2)}$,
USING NEUTRAL ATMOSPHERE MODEL WITH NO SEASONAL DEPENDENCE

PUERTO RICO

FIGURE 23



DRIFT VELOCITIES, PUERTO RICO

FIGURE 24

where the integration was stopped.

In general it was found that the drift and diffusion velocities were of the same order of magnitude, seldom exceeding about 8 meters/second, and showed a tendency to decrease during the night.

6. SUMMARY AND CONCLUSIONS

The purpose of this work was to carry out an investigation of the processes effective in the nighttime ionosphere and make estimates of the numerical values of the important coefficients.

This was done by integrating the continuity equation along a tube bounded by lines of the earth's magnetic field, and extending from the base of the ionosphere to the geomagnetic equator. In this way the loss processes and the drift and diffusion velocities were all taken into account.

6.1 Assumptions in the Analysis

Several assumptions were necessary in the recombination coefficient analysis and the effect of each of these was examined.

6.1.1 Top Profile

In order to integrate the continuity equation between the base of the ionosphere and the geomagnetic equator a top profile must be added to bottomside electron density profile. Three different top profiles were assumed and it was found that the variation in β_{300} due to the top profile assumption did not exceed 20 percent over the solar cycle.

6.1.2 Quadratic Loss Coefficient

The quadratic loss coefficient, α, must be included in the loss process for the lower altitudes. A range of values from

0.5×10^{-8} to 10^{-7} was assumed for α, and the effect on the computed value of β_{300} examined. For a temperature of 900°K, β_{300} increases by 0.13×10^{-4} sec^{-1} when α is changed from 10^{-8} to 0.5×10^{-8}; while at 1250°K for the same α change, the increase in β_{300} is about 0.18×10^{-4} sec^{-1}.

Values of the quadratic loss coefficient, α, between $0.2 \times 10^{-8} \times 1.0 \times 10^{-8}$ provided best agreement between the total and diffusion velocities profiles. It should be pointed out, however, that the assumptions about α and about the low-lying ionization are extremely interdependent and these limits for α should be regarded with some reservation.

6.1.3 Different Assumptions about the Molecular Constituent

It was assumed in the analysis that the linear loss coefficient varied in height in the same manner as either the O_2 or N_2 molecular density. It was found that the assumption that the effect of N_2 is predominant increases the average value of β_{300} calculated for all seasons by about 10 percent at 1200°K and by about 90 percent at 740°K over values calculated assuming O_2. In view of the small values of β_{300} at low temperatures in winter, the assumption that the effect of N_2 is predominant is attractive.

6.1.4 Low-Lying Ionization

The electron density profile data required a correction for

low-lying ionization which was neglected in the ionogram reduction method. When this correction was omitted the average value for β_{300} for all seasons increased by about 30 percent at 1240°K and about 75 percent at 750°K for a value of α of $.5 \times 10^{-8}$ cm^3 sec^{-1}. This correction is not very reliable and may well have been overestimated.

6.1.5 Production Due to Protonosphere Content Change

If a high estimate of 30 percent of the protonosphere is assumed to produce atomic oxygen ions by charge exchange, a change in β_{300} of 5 percent at 750°K and a change of 17 percent at 1250°K are noted. The change referred to here is again relative to the mean β_{300} at that temperature.

Because the winter values at low temperatures are so small and this effect is additive much larger percentage increases in β_{300} are observed in these cases.

6.1.6 Decrease in Te/Ti During the Night

If it is assumed that Te/Ti is 1.5 at 2300 hours and decreases to 1.0 during a five hour period, the change in β_{300} is about 20 percent at 750°K and about 25 percent at 1250°K. This change is also referred to the all season mean value of β_{300} and again the percentage increase for low sunspot, winter conditions is much larger.

6.2 Recombination Coefficient

6.2.1 The Linear Recombination Coefficient at 300 Kilometers

The recombination coefficient at 300 kilometers, β_{300} was calculated for 6 stations located between 29° and 59° North geomagnetic latitude. Much the same results were obtained for all stations and no latitude dependence was noted. The method breaks down for summer months for high latitude stations because the upper layer is sunlit throughout the night.

Data for 143 months for 6 stations were used in the analysis.

It was verified that a large variation in β_{300} over the solar cycle does occur.

Based on the results of this analysis the following best estimates for β_{300} are suggested:

Table 3

Suggested Summer and Winter Values for β

	Summer	Winter
T °K	β	β
1250	$1.95\times10^{-4}\exp-\left[\frac{Z-300}{35}\right]\sec^{-1}$	$0.90\times10^{-4}\exp-\left[\frac{Z-300}{35}\right]\sec^{-1}$
1025	$1.11\times10^{-4}\exp-\left[\frac{Z-300}{27}\right]\sec^{-1}$	$0.35\times10^{-4}\exp-\left[\frac{Z-300}{27}\right]\sec^{-1}$
750	$0.21\times10^{-4}\exp-\left[\frac{Z-300}{20}\right]\sec^{-1}$	$0.65\times10^{-6}\exp-\left[\frac{Z-300}{20}\right]\sec^{-1}$

6.2.2 Seasonal Effects

A large seasonal variation for β_{300} was observed. By independently determining the temperature from the 10.7 cm solar flux, and from the profile shape, it was determined that this seasonal variation is not merely a temperature effect.

The recombination coefficient was assumed proportional by a factor γ to the N_2 or O_2 densities. Since presumably γ is not dependent on season, the molecular densities O_2 or N_2 must have a large seasonal variation. Low molecular densities in winter at a given height and high moledular densities in summer would explain the winter anomaly in the daytime ionosphere as satellite density measurements show that the neutral atomic oxygen density, to which the production is proportional, is not very seasonally variable.

6.3 The Diffusion Coefficient

The diffusion coefficient for ions in the upper F region was determined using data for 84 individual hours. The normalized diffusion coefficient Dn(0) may be assumed proportional to T^k. When this was done an estimate of k of (0.8 ± 0.4) was made. Using a value for k of 0.8 the diffusion coefficient is given by:

$$D = \frac{(6.5 \pm 3) \times 10^{15} (T)^{0.8}}{n(0)} \quad cm^2 \; sec^{-1}$$

If it is assumed, following Dalgarno, that k = 0.5, then the diffusion coefficient is given by:

$$D = \frac{(0.5 \pm 0.3) \times 10^{17} \sqrt{T}}{n(0)} \quad cm^2 \, sec^{-1}$$

Because of the height range used this coefficient refers to 0^+ ions diffusing in atomic oxygen.

6.4 Drift Velocities

The drift velocity during the night was determined as part of the diffusion coefficient analysis. It was found that the drift velocities seldom exceeded 8 meters/second, and a trend of the drift velocity decreasing during the night was noted. No seasonal dependence was observed.

6.5 Suggestions for Further Research

This analysis of F region processes was based upon available ionospheric profiles and neutral atmospheric models.

The ionization density profiles required correction both above and below the peak of the layer. An extension of this work should be carried out using more accurate estimates of the true shape of the layer and its changes with time. Accurate profiles both above and below the maximum should allow a better estimate of the important coefficients to be made.

ACKNOWLEDGEMENTS

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I am grateful to Dr. Marcel Nicolet for several helpful discussions during the course of this work and for the neutral atmospheric models which he provided.

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